

**The Seiland Igneous Province, Northern Norway:
Age, Provenance, and Tectonic Significance**

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the requirements for the degree of
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Declaration

I declare that this dissertation is my own, unaided work. It is being submitted for the Degree of Doctor of Philosophy in the University of the Witwatersrand, Johannesburg. It has not been submitted before for any degree or examination in any other University

(Richard James Roberts)

On the 22nd day of March, 2007

Abstract

The Seiland Igneous Province, of which 5400 km² is exposed, is hosted within a discrete terrane within the northernmost part of the Caledonian orogenic belt. The Province consists of numerous mafic and ultramafic plutons emplaced into a sedimentary succession indicative of a continental setting. Accompanying this mafic magmatism is a significant volume of intermediate monzonitic and dioritic rock (10% of the total exposed igneous rock), as well as numerous nepheline syenite and carbonatitic intrusions.

This study reports ID-TIMS U-Pb analyses on magmatic zircons from a range of intrusions, which indicate that the bulk of the Seiland magmatism took place between 560 Ma and 570 Ma, whereas previous studies had produced a range of ages between 420 Ma and 830 Ma. The data indicate that only one magmatic episode is represented in the rocks of the Seiland Igneous Province, invalidating previous models involving multiple rifting events over a period of 300 m.y.

Detailed geochemical investigation of several plutons from an evolved high alkali suite of gabbroic intrusions in the Seiland Igneous Province has shown that these plutons are generally enriched in trace elements compared to layered intrusions from other areas across the globe, but that geochemically the gabbros are relatively homogenous. The rocks yield ϵ_{Hf} and ϵ_{Nd} values for the gabbroic rocks ranging from +8 to -6 and from +4 to -4, respectively, indicative of the contamination of mantle-derived material with crustal material. The most primitive isotopic values are similar to those obtained from the carbonatites and nepheline syenites, indicating the same mantle source gave rise to the magmas that were subsequently emplaced as the Seiland Igneous Province. The homogeneous trace element content of the different mafic rocks most likely indicates a relatively homogeneous mantle source for the original magmas of the province, which has subsequently been affected by processes of assimilation and crustal contamination. The monzonitic and dioritic bodies in the Seiland Igneous Province are not derived from melted silicic crustal material and may have been formed by the melting of pre-existing mafic material.

The new geochronology invalidates the metamorphic framework previously proposed for the Seiland Igneous Province, which postulated several orogenic events between the emplacement of the magmas and the Caledonian Orogeny. There is no evidence for metamorphic activity in the period between 570 Ma and 420 Ma, and there are monazites in gneissic rocks hosted within mafic rocks of Seiland age that preserve an age of 640 Ma. This leads to the conclusion that only one metamorphic event, the 420 Ma Caledonian Orogeny caused by the collision of Baltica and Laurentia, affected the Seiland terrane after the emplacement of the Seiland magmas.

The new data obtained lead to a model for the evolution of the Seiland Province in which a number of heavily modified and contaminated mantle-derived mafic magmas derived from the mantle were emplaced into the continental crust of the Seiland nappe between 560 and 570 Ma. This magmatism was accompanied by the injection of alkaline magmas into the same area of the crust, and the melting of mafic rock emplaced earlier. This magmatic event is considered to have occurred in an extensional stress regime, possibly during intracontinental rifting or back-arc spreading. This event took place well before the 420 Ma Caledonian Orogeny, and thus the Seiland Igneous Province can be considered a remnant of an older geological terrane that was emplaced onto the margin of Baltica during the Caledonian Orogeny.

To all those who have inspired me over the years

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Chapter 1: Introduction

The geological record on this planet is often obscured by the very processes we, as geologists, wish to investigate. Continents rift apart, rotate and collide, again and again, and the destructive processes involved in subduction and continental collision often destroy the evidence for the past history of the rocks involved. In northern Europe, the mountainous Caledonides of Palaeozoic age effectively mask a large portion of Precambrian Earth history related to the craton of Baltica, as much of this older rock has been metamorphosed, buried or destroyed during the orogenic event. However, there are much older rocks within this orogenic belt, most of which are only coming to light now.

This thesis reports the results of a project focused on unravelling the geological history of a suite of igneous rocks in northern Norway. This suite of rocks, referred to as the Seiland Igneous Province (SIP) after the central island in the region (Figure 1.1), has presented a number of problems in deciphering the tectonic history of this portion of the Norwegian coast. Despite having been studied for more than fifty years (e.g. Barth 1953; Heier 1960, 1961, 1964), there has been no consensus on the age, tectonic setting or post-emplacement history of the SIP (e.g. Ramsay et al., 1985; Daly et al., 1991), and, as such, the SIP is an ideal candidate for the application of modern geochronological and geochemical techniques.

1.1. The geology of the Seiland area and its surroundings

The fold and thrust belt of the Scandinavian Caledonides is by far the defining feature of Norwegian geology. This thrust belt, created during a collision between the Laurentian and Baltican Cratons during the Scandian (420 Ma) era, stretches from the north-east coast of North America, across the northern part of the British Isles, and along the length of Norway and northern Sweden. Despite the deformation accompanying the orogenic event creating the Caledonides, it is still possible to identify portions of pre-Caledonian terranes amongst the nappes thrust up onto the craton of Baltica.

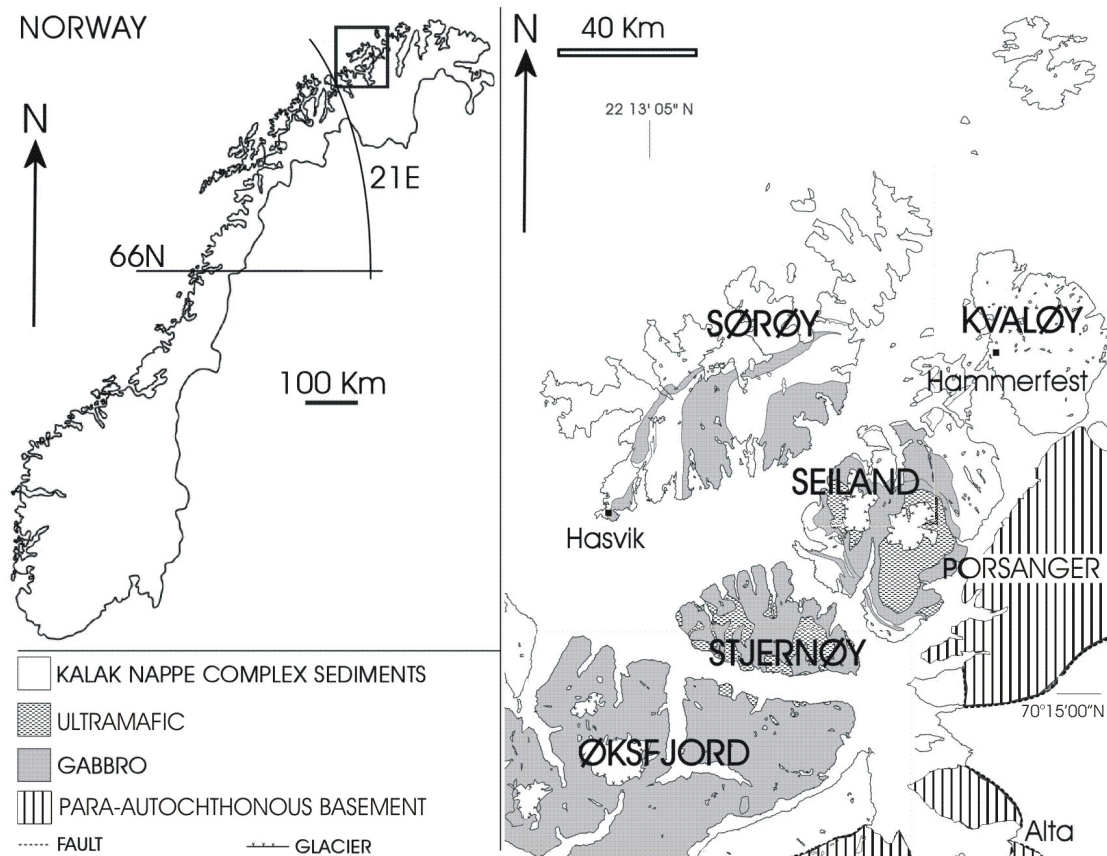


Figure 1.1- The map of Norway on the left shows the location of the western Finnmark region. The detailed map of the area outlined in the regional map is on the right, showing the distribution of the major igneous bodies within the Seiland Igneous Complex.

In Finnmark, northern Norway, Caledonide geology was originally held to comprise an Eocambrian to Cambrian sequence, thrust over a Precambrian basement and its unconformable cover of late Proterozoic to Ordovician sediments (Sturt et al., 1978; Ramsay et al., 1985). The overthrust package consists of a major upper nappe, the Kalak Nappe Complex, resting on smaller and thinner parautochthonous nappes (Holtedahl & Dons, 1960). In general, each nappe of the complex is characterized by a basal plinth of Precambrian rocks, which can be banded orthogneiss, paragneiss or some combination of both (Sturt, & Ramsay, 1978; Ramsay et al., 1985). This plinth is covered by an unconformable metasedimentary cover sequence. A type succession for regional correlation between the members of the Kalak Nappe Complex is provided by the lithostratigraphy of the Sørøy Group on the island of Sørøy (Figure 1.1). The basal member of the Sørøy Group, the Klubben Psammite, is widely distributed in Finnmark, whereas the younger sequence comprising the Storelv

Formation schists, Falkenes Formation marbles and Hellefjord Formation schists is more limited in extent.

The dominant features of the Sørøy Nappe, west of Kvaløy, are the igneous rocks of the SIP shown in Figure 1. This complex crops out over an area of c. 7000 km² and comprises numerous plutons, ranging in composition from calc-alkalic to tholeiitic and alkaline gabbro (Robins & Gardner, 1975), along with large ultramafic plutons (Bennet et al., 1986) and monzonitic and dioritic sheets (Robins & Gardner, 1975). Nepheline syenite (Robins, 1972), carbonatite (Sturt & Ramsay, 1965) and granite (Sturt & Taylor, 1971) are also present as dykes, stockworks and small discrete intrusions. At some point after the emplacement of the plutons, a regional event metamorphosed some of the rocks to amphibolite facies, locally producing deformation fabrics such as foliations (e.g., Sturt & Ramsay, 1965).

Some previous studies have focused on the petrological examination of the different plutons within the SIP, but there has been no consensus as to the age or origin of the province as a whole. The SIP has been considered to be related to Andean-type magmatism in a subduction setting (Robins & Gardner, 1975), a hot spot (Stephens et al., 1989), and a centre of extensional rifting (Krill & Zwaan, 1987; Daly et al., 1991). However, the greatest obstacle to understanding the SIP has been the poor age constraints on the province, which has led to significant confusion.

1.2. Previous ideas on the tectonic history of the SIP

The Finnmark region of northern Norway has long been held to show the effects of an orogenic event prior to the ca. 420 Ma Scandian phase of the Caledonian Orogeny, the tectonic episode responsible for the transport and emplacement of nappes on the craton of Baltica (Sturt et al., 1975, 1978). This “Finnmarkian” event (Sturt et al., 1975, 1978), originally considered a phase of the Caledonian Orogeny and later upgraded to a separate orogeny, was defined based on the field relationships between the voluminous intrusions of the Seiland area and the Kalak Nappe Complex that hosts the intrusions. The intrusive rocks crosscut pre-existing fabrics, and were themselves subsequently deformed along with the Kalak Nappe Complex.

Early work in the Finnmark region identified only two phases of deformation in the Kalak Nappe Complex, denoted D1 and D2 (Sturt & Ramsay, 1965). However, later work recognized an additional earlier schistosity in rocks of the Sørøy Group (Sturt, Pringle & Roberts, 1975; Sturt, Pringle & Ramsay, 1978). Consequently, the most recently postulated structural framework for the region involves three episodes of deformation (D1-D3), of which D2 and D3 are by far the most prominent (Ramsay et al., 1985). Thus, the Hasvik Gabbro originally described as “post-D1” in age (Sturt & Taylor, 1971), is now described as “post-D2” by Sturt, Pringle & Ramsay (1978). In the ensuing sections, the D1-D3 deformation scheme of Sturt, Pringle & Ramsay (1978) is used.

The earliest deformation event (D1) is poorly preserved in Finnmark. The evidence for its existence comes mainly from a widespread schistosity developed in pelitic and semi-pelitic lithologies of the Sørøy Group. The second deformation event (D2) is by far the most dominant one in Finnmark (Sturt & Ramsay, 1965; Sturt, Pringle & Roberts, 1975; Sturt, Pringle & Ramsay, 1978; Ramsay et al., 1985). Recumbent folds, with large (km) amplitudes and wavelengths, and an axial planar foliation herald an event that was associated with amphibolite facies metamorphism. The subsequent D3 event, the last significant episode of deformation, involved a lower metamorphic grade (greenschist facies), and can be seen in the refolding and reorientation of D2 structures and the development of a weak axial plane cleavage.

Assessing these three deformational events and placing them in their correct spatial and temporal context has long been a problem for geologists working in the area. Before the identification of ages older than 420 Ma in the rocks of the area, the D2 event was considered to represent an early stage of the Caledonian Orogeny (e.g., Sturt & Ramsay, 1965). As older ages emerged and a time gap between D2 and D3 was identified, a separate orogeny, the Finnmarkian (490-540 Ma), was proposed as the source of D2 deformation (Sturt, Pringle & Roberts, 1975; Sturt, Pringle & Ramsay, 1978; Ramsay et al., 1985). The D3 deformation was then tentatively correlated with Scandian (Caledonian) deformation at 420 Ma (Ramsay et al., 1985).

These deformational structures were originally correlated across much of Finnmark, creating numerous problems of interpretation. It is now known (Daly et al., 1991) that

comparable structures developed at different locations during totally unrelated events ranging in age from the Precambrian to the Silurian. The inconsistent interpretations led to various debates and to the formulation of alternative models (e.g., Binns 1989). Krill and Zwaan (1987) questioned the original structural interpretation by Ramsay et al. (1985) and postulated that the SIP formed in an extensional environment. However, their contention that there were no undeformed mafic dykes in the area was widely contested (e.g., Sturt and Ramsay 1988) and can be readily disproved in the field, where numerous undeformed dykes can be observed. Daly et al. (1991) used a variety of isotopic methods (Rb/Sr, Sm/Nd, U/Pb) to date various igneous intrusions and demonstrated that the “D2” deformation of the Klubben Psammite on the Porsanger peninsula took place before 800 Ma. The model resulting from their work questioned the existence of the Finnmarkian Orogeny, and proposed a 250 m.y. period in the Precambrian during which rifting took place. This idea was further developed by Elvevold et al. (1994), who proposed a three-stage metamorphic history reflecting early rifting and magmatism followed by crustal thickening and contraction. Such a long period of rifting seems unusual, and would imply the opening of a sizeable ocean basin or multiple episodes of rifting.

1.3. Isotopic dating in the Seiland area

The dating of the Finnmarkian Orogeny has always been contentious. Initial palaeontological evidence (Holland & Sturt, 1970) implying an Early Cambrian age for calc-silicate rocks in the Sørøy Group has been disproved (Debrenne, 1984), but early K-Ar and Rb-Sr studies from Sørøy indicated an age of 540- 490 Ma for mafic rocks (Sturt, Miller & Fitch, 1967; Sturt, Pringle & Ramsay, 1978). A Rb/Sr age of 612 ± 17^1 Ma was reported for a suite of syenite-monzonite-gabbro and peridotite from the Øksfjord peninsula (Brueckner, 1973), but was not widely accepted. The occurrence of a metamorphic event at ca. 490 Ma, as well as the later Scandian event (420-400 Ma), was indicated by Ar-Ar geochronological studies in northern Norway (Dallmeyer, 1988). The wide spread of ages was used in constructing a petrological framework for the emplacement of the different gabbros by Robins and Gardner

¹ All Rb-Sr ages calculated using $\lambda^{87}\text{Rb} = 1.42 \cdot 10^{-11} \text{ a}^{-1}$

(1975), who argued for an Andean margin-style tectonic setting, with gabbroic intrusions evolving from tholeiitic to calc-alkaline compositions with time.

More recent Sm-Nd geochronology resulted in a complicated scenario for the evolution of the Finnmark region. In a paper incorporating numerous studies on the area, Daly et al. (1991) provided ages for different intrusions in western Finnmark spanning a period of 250 m.y. from 851 ± 130 Ma to 604 ± 44 Ma. Other workers obtained intrusion ages in the same range, using both Rb-Sr and Sm-Nd techniques (Mørk & Stabel, 1990; Krogh & Elvevold, 1991). Not only were some of these ages significantly older than all previous dates from the area, they also implied episodic magmatism over a very long period. Determining the tectonic setting capable of producing basic magmatism over such a long period has proved difficult, and has resulted in complex models involving multiple rifting events and several ocean closures (e.g., Reginiussen, 1996).

Only three igneous bodies in the Finnmark area were dated by the U-Pb zircon isotope dilution method prior to the current study. The Litlefjord Granite on the Porsanger peninsula, emplaced into the Klubben Psammite of the Sørøy Group, was dated at 804 ± 19 Ma using a lower Concordia intercept age (Daly et al., 1991). On the basis of this result, a new Late Proterozoic, “Porsanger” Orogeny was proposed, which was tentatively placed at 850-800 Ma or older on the basis of the age from Litlefjord and the Rb/Sr age of 851 ± 130 Ma from a granitoid at Repvåg (Daly et al., 1991). Nepheline syenite pegmatites that intrude mafic to ultramafic complexes on the islands of Seiland and Stjernøy were dated by U/Pb at 531 ± 2 and 523 ± 2 Ma, respectively (Pedersen et al., 1989). These pegmatites were argued to be the latest intrusions in the SIP, as they crosscut all other intrusive bodies in the sampling areas and are considered post-tectonic, and these ages have therefore been interpreted to represent a lower limit on deformation in the area.

1.4. Aims and structure of this thesis

This project is focussed on re-examining various aspects of the SIP in a regional context, and reports the results of both fieldwork and analytical investigation carried out over three field seasons in northern Norway. The project was supervised by

Professors L.D. Ashwal (University of the Witwatersrand) and T.H. Torsvik (Geological Survey of Norway). Dr. Donald Ramsay served as a field guide to the Finnmark region of Norway. Chapter 2 is an overview of the petrology and field relationships for the different intrusive bodies of the SIP, compiled from previous work and from field observations in the area over the course of three field seasons. This review is of great importance in providing a background for subsequent chapters, and placing subsequent analytical work and observations within a firm geological framework.

Of primary importance in understanding the relationships between the igneous rocks of the SIP and their country rocks is an accurate chronological framework, and so a large portion of the analytical work undertaken during this project has been focussed on obtaining high precision ages from the different intrusive bodies within the SIP. This dating focussed on U/Pb ID-TIMS analysis of zircon, monazite and titanite, and was conducted under the supervision of Prof. Fernando Corfu at the University of Oslo, Norway. Chapter 3 presents the analytical methodology used in subsequent chapters. Chapter 4 details the results of the dating of several different mafic, felsic, and alkaline intrusions within the SIP, and has been taken from the following papers,:

Roberts, R.J., Corfu, F., Torsvik, T.H., Ashwal, L.D., & Ramsay, D.M., 2006, Short-lived mafic magmatism at 560-570 Ma in the northern Norwegian Caledonides - U/Pb zircon ages from the Seiland Igneous Province, *Geological Magazine*,

Roberts, R.J., Corfu, F., Torsvik, T.H., Hetherington, C.J., and Ashwal, L.D, Alkaline and carbonatitic magmatism coeval with mafic plutonism in the Seiland Igneous Province, Northern Norway: age and palaeotectonic significance, submitted to the *Journal of the Geological Society*, London.

Dr. C.J. Hetherington provided SEM images of problematic zircons and an interpretation of the zircon textures revealed in the images, as well as numerous observations during one season of fieldwork in northern Norway.

Once the age relationships between the different components have been analysed, it is possible to start investigating the geodynamic and petrological issues related to the SIP. Magmatism does not occur randomly on this planet, but is often intimately related to the geodynamics of plate motion and mantle convection. In this study, the identification of the tectonic setting in which the SIP was emplaced, and the source of the magmas produced in such a setting are two important concerns.

Chapter 5 provides details on the geochemical evolution of the SIP. Both whole rock and trace element geochemistry are reported, and used in conjunction with the geochronological framework formulated in previous chapters to investigate various aspects of the SIP's evolution. Chapter 5 also reports the results of Hf isotopic analyses of zircons from the different SIP intrusions, as well as whole rock Sm/Nd isotopic analyses of several different rocks. These data are used to constrain the mantle source of some of the magmas of the SIP, and to investigate their subsequent evolution during emplacement. Chapter 6 reviews the metamorphic studies previously conducted in the SIP in the light of the evidence collected in previous chapters and evidence from a metamorphic paragneiss in the Øksfjord area. Finally, Chapter 7 reports the conclusions reached in this study.

Chapter 2: The igneous intrusions of the Seiland Igneous Province

The literature on the intrusions of the Seiland Igneous Province (SIP) stretches back for more than fifty years. This chapter summarises much of the previous work on the different intrusions comprising the province, and reports observations made in the field and on thin sections collected during the course of this study. This will provide a suitable background against which new analytical data can be presented and discussed. The different categories of intrusion within the complex are discussed first in general, and then specific intrusions are detailed according to their spatial distribution. Modal mineralogies for listed samples are reported in Table 2.1., along with the GPS co-ordinates for the localities where they were sampled. Photomicrographs of the majority of the samples are presented in Appendix A.

2.1. The mafic plutons

The mafic plutons of the SIP are the largest component of the province, and are commonly the hosts for later ultramafic and alkaline intrusions. Robins & Gardner (1974) divided the plutons into three groups on the basis of mineralogy and limited major element chemistry:

- 1) Gabbroic intrusions with a tholeiitic parental magma. These intrusions typically show stratiform layering.
- 2) “Syenogabbro”- an igneous complex composed of interlayered gabbroic rock and more evolved igneous rocks, typically syenite, monzonite and diorite, considered by Robin & Gardner (1974) to represent silica- saturated derivatives of the mafic magmas.

3) Clinopyroxene gabbros. These gabbros are always nepheline normative, and contain alumina-rich clinopyroxene and calcic plagioclase.

Robins & Gardner (1974) consider the first two types of intrusion to be derived from similar magmas that underwent different processes of differentiation, either undergoing enrichment in iron or becoming silica-oversaturated (or both). The third type of intrusion was held to be derived from a different parental magma, of alkaline olivine basalt composition (Robins & Gardner, 1974). Under the geochronological framework accepted at the time, each of the different types of mafic pluton was emplaced in distinct time periods, with tholeiitic gabbros being both the oldest and youngest plutons in the area, and the “syenogabbro” and clinopyroxene gabbros being emplaced between episodes of tholeiitic magmatism.

This scheme has a number of flaws. The definition of the different categories lack quantification, and the mineralogical characteristics in the scheme are vague and difficult to apply during field investigations. Furthermore, the nature of the magma parental to the liquid from which a plutonic igneous rock has crystallised is not directly represented in the mineralogical assemblage of that plutonic igneous rock, especially if the rock is a cumulate rock (e.g. O’Hara and Herzenberg, 2000).

Processes in magma chambers like magma mixing, assimilation of crustal material, and fractional crystallisation all work to obscure the original magma composition. Furthermore, cumulate rocks are not representative of the original magma composition, as crystals are no longer located stratigraphically where they crystallised.

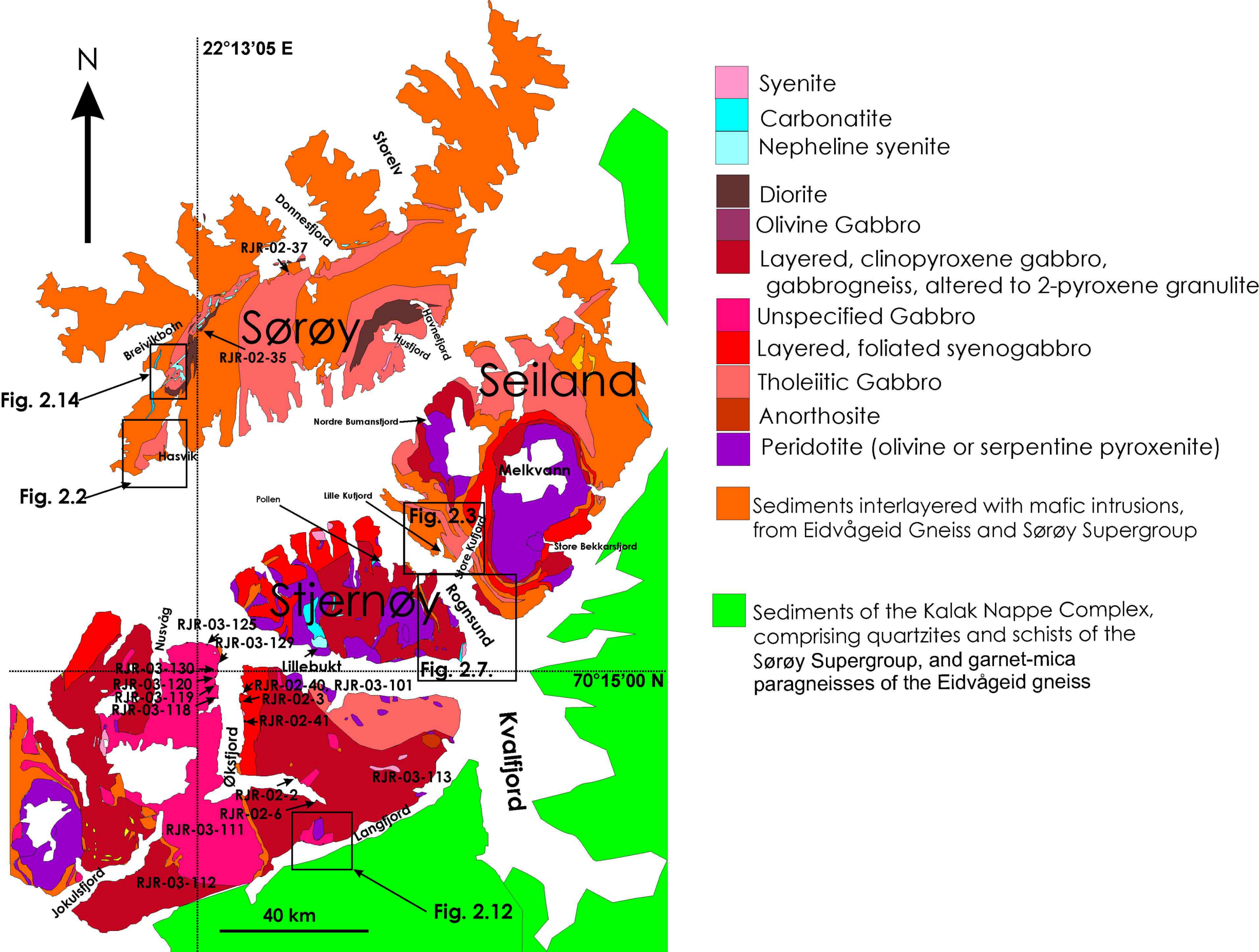


Figure 2.1. Colour map of the intrusions of the Seiland Igneous Province. Figures indicated refer to more detailed maps. Locations not detailed in the smaller maps are marked on the map. Taken from Roberts, D., 197, Geologisk kart over Norge, berggrunnskart Hammerfest 1:250 000, Norges Geologiske Undersøkelse [Bulletin 61, 1–49]. Legend for mafic rocks also taken unaltered from this source

Table 2.1. Sampling locations for this study, with GPS co-ordinates (WGS84) and visually estimated mineralogical modes for the rocks sampled

Sample	Location		Pl	Kfs	Ol	Opx	Cpx	Hbl	Bt	Qtz	Np	Ca	Op	Name	Notes
Hasvik Gabbro															
RJR-02-29I	N 70.29'03.9''	E 22.14'42.1''	50			20	25						5	Gabbro	Coarse-grained – Marginal Border Series
RJR-02-29J	N 70.29'03.9''	E 22.14'42.1''	50			20	25						5	Gabbro	Coarse-grained- Basal Zone
RJR-02-30A	N 70.29'04.1''	E 22.14'31.5''	40			5	5	40	5				5	Hornblende Metagabbro	Magnetite-opx symplectites- Upper Zone
RJR-02-30B	N 70.29'04.1''	E 22.14'31.5''	50			10	25	10	2				3	Gabbro Hornblende	Magnetite-opx symplectites- Upper Zone
RJR-02-30D	N 70.29'03.9''	E 22.14'30.1''	55				20	20		3			2	Metagabbro	Upper Border Series
Breivikbotn Gabbro															
RJR-02-31	N 70.32'56.2	E 22.14'42.1''	45				5	40						Hornblende Metagabbro	10% opx-green spinel symplectites
RJR-02-35	N 70.35'50.0''	E 22.21'08.7''	50	10				5	15	20				Tonalite	Quartz diorite
Storelv Gabbro															
RJR-02-37B	N 70.37'47.3''	E 22.34'30.0''	50			1	3	20	20				5	Hornblende Metagabbro	1% apatite
RJR-02-37A	N 70.37'47.3''	E 22.34'30.0''		50				20	20	10				Alkali Feldspar Granite	Zircon as accessory
Øksfjord- Southern end															
RJR-02-02	N 70.09'9.60''	E 22.25'28.9''	30				20	40					5	Hornblende Metagabbro	5% opx-green spinel symplectites
RJR-02-06A	N 70.06'51.1''	E 22.31'58.6''	30		5		30	30					5	Olivine Hornblende Metagabbro	Opx-green spinel symplectites visible
RJR-02-06B	N 70.06'51.1''	E 22.31'58.6''	50		5		10	10					15	Olivine Hornblende Metagabbro	10% green spinel
RJR-03-112	N 70.03.35.2"	E 21.54'15.8"	30			30		25					10	Garnet- Hornblende Metanorite	5% poikilitic garnet
RJR-03-113A	N 70.12'7.7"	E 22.35'20.0''	50		5	10	15							Olivine Hornblende Metagabbro	10% opx-green spinel symplectites
RJR-03-113B	N 70.12'7.7"	E 22.35'20.0''	50		10	10	15		5					Olivine Hornblende Metagabbro	10% opx-green spinel symplectites

Table 2.1. Sampling locations for this study, with GPS co-ordinates (WGS84) and visually estimated mineralogical modes for the rocks sampled

Sample	Location		Pl	Kfs	Ol	Opx	Cpx	Hbl	Bt	Qtz	Np	Ca	Op	Name	Notes
Øksfjord- Eastern side															
RJR-02-3D	N 70.13'26.6''	E 22.20'4.2''	20		2	1	35	40					1	Olivine Hornblende Metagabbro	Opx-green spinel symplectites visible
RJR-02-40B	N 70.13'52.0''	E 22.20'41.7''	60	35						5				Monzonite	Monzonite
RJR-02-41C	N 70.12'56.8''	E 22.20'06.0''	55	40						5				Monzonite Alkali feldspar syenite	Monzodiorite
RJR-03-101E	N 70.13'52.0''	E 22.20'41.7''		95			1		2				1		1% epidote
Øksfjord- Western side															
RJR-03-118	N 70.13'52.3"	E22.16'12.8"	50			20	25						5	Gabbronorite	
RJR-03-118B	N 70.13'52.3"	E22.16'12.8"	48			23	25		2				2	Gabbronorite	
RJR-03-120	N 70.15.28.2"	E22.16'34.0"	50			30		15					5	Hornblende Norite	
RJR-03-125	N 70.17'6.8"	E22.14'58.1"	30			30	40	3	2				5	Gabbronorite	Zircon and apatite as accessories
RJR-03-129A	N 70.17'15.2"	E22.10'27.3"	15	35						45			5	Granite	
RJR-03-129B	N 70.17'15.2"	E22.10'27.3"	20	22		10				45			3	Granite	Granite, with accessory garnet and zircon
RJR-03-129C	N 70.17'15.2"	E22.10'27.3"	50			30		15						Hornblende Norite	Granite, with accessory garnet and zircon
RJR-03-129D	N 70.17'15.2"	E22.10'27.3"	20			70			5				5	Orthopyroxenite	
RJR-03-130A	N 70.17'19.4"	E22.11'8.9"	35			10	10	30					10	Hornblende Gabbronorite	Apatite and zircon as accessories
Tappeluft Ultramafic Complex															
RJR-02-7	N 70.06'19.3''	E 22.31'10.0''	15		5		20	60						Hornblendite	
RJR-02-8	N 70.04'52.4''	E 22.26'50.6''		90					5				5	Syenite	Magnetite-spinel- opx-cpx symplectites
RJR-03-108A	N 70.06'19.3''	E 22.31'10.0''	10		40	10		30					5	Harzburgite	5% magnetite-green spinel- opx symplectites
RJR-03-109	N 70.04'54.8"	E 22.28'14.1	40					40	10				10	Hornblende Gabbro	Pegmatitic gabbro

Table 2.1. Sampling locations for this study, with GPS co-ordinates (WGS84) and visually estimated mineralogical modes for the rocks sampled

Sample	Location	Pl	Kfs	Ol	Opx	Cpx	Hbl	Bt	Qtz	Np	Ca	Op	Name	Notes
Breivikbotn Carbonatite Complex														
RJR-02-34D	N 70.33'58.5''	E 22.15'06.7''				20		5			40		Silico-carbonatite	20% talc, 4% garnet, 1% apatite
RJR-04-224	N 70.33'58.5''	E 22.15'06.7''			2	3	2				20	3	Scapolite rock Garnet	65% scapolite, 5% apatite
RJR-04-225	N 70.33'58.5''	E 22.15'06.7''			7.5	5					45	5	silicocarbonatite	12.5 % garnet, 30% scapolite
RJR-04-226	N 70.33'58.5''	E 22.15'06.7''	70								5		Scapolite diorite Garnet	25% scapolite
RJR-04-227	N 70.33'58.5''	E 22.15'06.7''			7.5	5					45	5	silicocarbonatite	12.5 % garnet, 30% scapolite
RJR-04-228	N 70.33'58.5''	E 22.15'06.7''			20	5					60	10	Carbonatite	5% apatite
Breivikbotn Alkaline Complex														
RJR-03-115	N 70.33'17.8"	E 22.15'52.7"	20	40			5	10		20		3	Nepheline Monzosyenite	2% muscovite
RJR-03-116	N 70.33'46.2"	E 22.16'0.2"	20	30			5			30		3	Nepheline Monzosyenite	10% muscovite, 2% epidote
RJR-04-236	N 70.33'17.8"	E 22.15'52.7"	20	40			5	10		20		3	Nepheline Monzosyenite	2% muscovite
RJR-04-245	N 70.32'45.4"	E 22.15'10.8''		90				2		4		4	Nepheline-bearing syenite	
RJR-04-246	N 70.32'45.4"	E 22.15'10.8''	25	40				8		25			Nepheline Monzosyenite	2% muscovite, zircon as accessory

Notes: Pl = Plagioclase Hbl = Hornblende
Kfs = K-feldspar Bt = Biotite
Ol = Olivine Qtz = Quartz
Opx = Orthopyroxene Np = Nepheline
Cpx = Clinopyroxene Ca = Calcite
Op = Opaques

There are several other flaws present in the classification. The clinopyroxene gabbro type proposed by Robins & Gardner (1974) lacks definition, in that no geochemistry or mineral analyses support the proposed mineralogy, and the idea of a primary alkaline olivine basalt parental melt is not supported in the literature (e.g. O'Hara & Herzenberg, 2002). The third member of the classification scheme, the “syenogabbros” also present problems, in that the process of mafic melt crystallisation makes it difficult to derive any significant amount of felsic melt from a mafic melt (e.g. Bowen, 1928; Longhi, 1991).

2.1.1. The Hasvik Intrusion, Sørøy

The Hasvik Intrusion (Fig. 2.2) lies at the southwestern tip of the island of Sørøy. A submarine aeromagnetic anomaly covering an area of 30 km² off the coast of the island indicates that part of the pluton is submerged, but an area of 12 km² is well exposed (Tegner et al., 1999). The intrusion is emplaced into the overturned limb of an antiform in the country rocks of the Klubben Quartzite, and is itself deformed into an asymmetrical syncline with a SW-plunging axis (Tegner et al., 1999).

The Hasvik Intrusion is a layered pluton displaying well-preserved cumulus textures. It was originally described as a “tholeiitic” gabbro by Robins & Gardner (1975). The pluton is subdivided into a Marginal Border Series, extending from the wall of the magma chamber, a Layered Series, extending from the floor of the chamber, and an Upper Border Series, extending from the roof of the chamber down (Robins & Gardner 1974). The Marginal Border Series comprises a thin layer of gabbro containing abundant crustal xenoliths of quartzitic composition, followed by a coarsening-inwards sequence of homogenous olivine gabbro, for a total thickness of up to 100 m

(Tegner et al., 1999). The Upper Border Series is poorly preserved, and outcrops as a thin, 60 m sheet of massive, Fe-oxide bearing gabbronorite at the highest elevation in the pluton (Tegner et al., 1999).

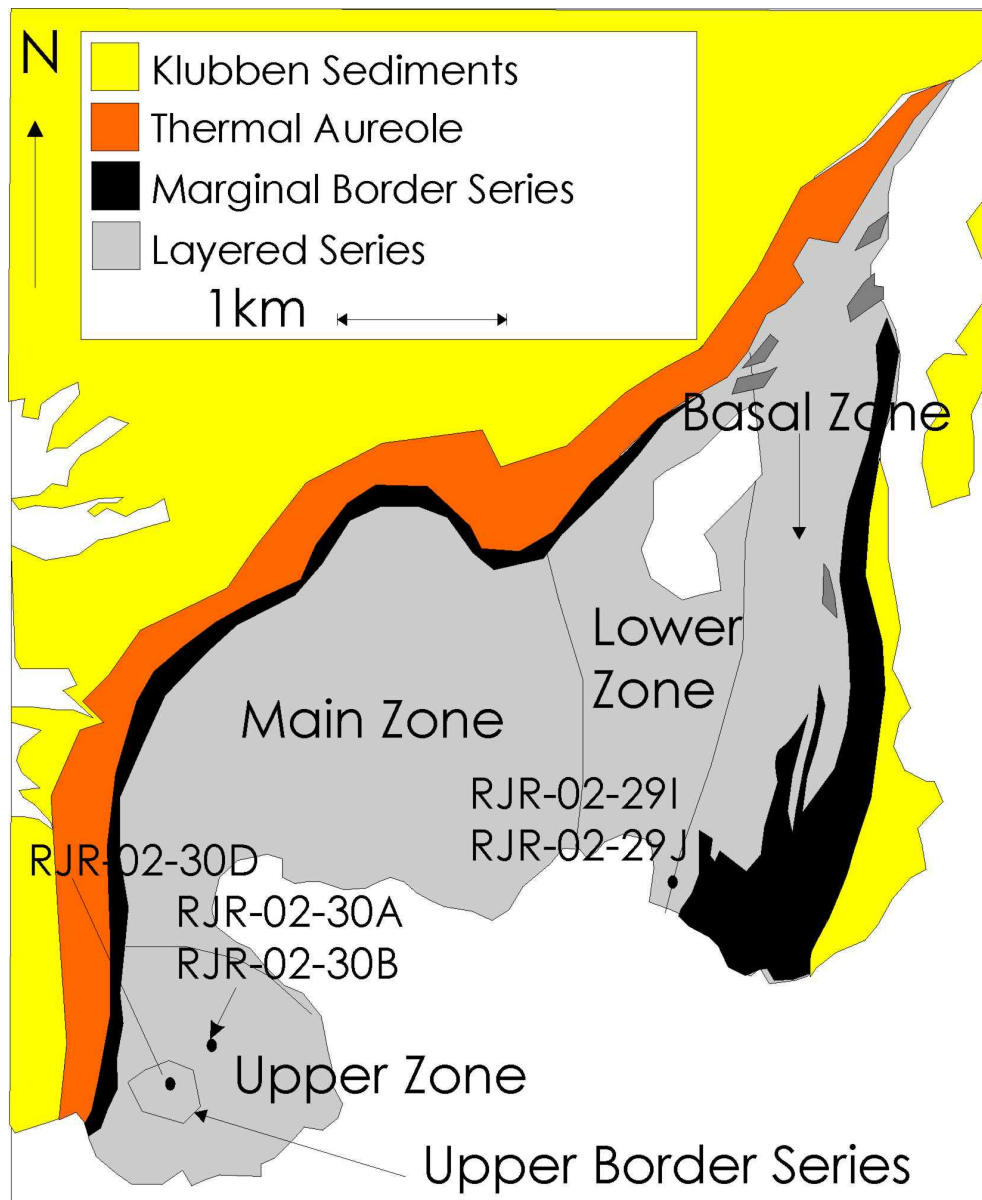


Figure 2.2- The Hasvik Intrusion, after Tegner et al. (1999).

The Layered Series comprises at its thickest a ± 1550 m thick cumulate sequence.

Stratigraphically, the Basal Zone is the lowest member of the sequence, comprising ± 75 m of laminated gabbronorite, much of which is interfingered with the country

rocks of the pluton's contact aureole. This Basal Zone is heavily contaminated with quartzitic crustal material, and contains numerous micro-xenoliths of the local Klubben Quartzite (Tegner et al., 1999). Mineralogically, the Basal Zone is olivine-free, and comprises plagioclase, orthopyroxene and augite). The succeeding Lower Zone (± 335 m) comprises olivine gabbro with distinct layering, and is the most primitive part of the intrusion (whole rock Mg-numbers for this sequence range from 0.688- 0.72; Tegner et al., 1999). The mineral assemblage contains plagioclase, augite and olivine, but orthopyroxene is only occasionally present. The Main Zone (± 950 m) is marked by the occurrence of Fe-Ti oxides (magnetite and ilmenite; Robins 1985) and orthopyroxene, and the disappearance of olivine. The uppermost member of the Layered Series is the ± 190 m thick Upper Zone, marked by the appearance of apatite, pigeonite (inverted to orthopyroxene with augite lamellae during cooling), and intercumulus quartz and quartz-feldspar intergrowths (Tegner et al., 1999). Major element and some trace element analyses, along with Sm/Nd and Rb/Sr isotopic ratios, from all the rocks of the Hasvik Intrusion have been kindly supplied by Christian Tegner (*pers. comm.*), and are located in Appendix A.

The Hasvik Intrusion is held to have crystallised by assimilation-fractional crystallisation processes (AFC) from a contaminated tholeiitic melt (Tegner et al., 1999). This can clearly be seen in that the Lower Zone is more primitive than the Basal Zone of the intrusion, and that AFC modelling accurately predicts the subsequent evolution of the Layered Series (Tegner et al., 1999). The contaminant involved in the evolution of the pluton is held to be the impure Klubben Quartzite country rock local to the pluton, and an estimate of the degree of contamination present from Nd and Sr isotopic analysis of the pluton indicates that 21 % in bulk of

the pluton originated from assimilated crust (Tegner et al., 1999; Tegner et al., 2005). However, the crystallisation sequence followed by the cumulate series in the Hasvik Intrusion is broadly similar to that in the Skaergaard Intrusion (Wager & Brown, 1968), except in the final stages of crystallisation. At Skaergaard, the most evolved cumulates are iron-rich, whereas the Upper Zone of the Hasvik Intrusion is silica-rich (Tegner et al., 1999).

The intrusion has a well-defined contact metamorphic aureole of at least 500 m that displays evidence of dehydration and partial melting, and the growth of garnet and hercynitic spinel (Reginiussen, 1996). Estimates obtained by geothermobarometry on garnet-orthopyroxene-plagioclase-quartz equilibrium (Bohlen et al., 1983) and aluminium solubility in orthopyroxene (Harley, 1984) indicate an emplacement pressure of 6- 8 kbar (Reginiussen, 1996). However, the interpretation of structural relationships in the aureole and, hence, the timing of the intrusion of Hasvik Gabbro in the deformational framework of the Finnmark region, has been contentious. The earliest work on the Hasvik Gabbro considered the gabbro to have been emplaced within the earliest phase of deformation within the nappes (Sturt, 1969). The contact aureole was observed to post-date one phase of folding and, in the two-stage structural framework of the time, was assigned to a post-D1 age. It can be readily observed that the country rock xenoliths preserved within the gabbro display isoclinal folding, indicating that folding had already occurred in the country rocks at the time of intrusion. This post-D1 designation was carried through the literature until Daly et al. (1991) described two phases of deformation within the aureole of the intrusion, and assigned the Hasvik intrusion to a post-D2 event. The evidence provided was that the contact of the intrusion truncates both the S1 and the S2 foliations (according to the

newer scheme of deformation; Ramsay et al., 1985) in the country rocks. Subsequent observations by Reginiussen (1996) and Tegner et al. (1999) support this interpretation. It is now also evident that the Klubben Quartzite underwent deformation and metamorphism during orogenic events predating the SIP (Daly et al., 1991; Corfu et al., 2006).

A variety of ages have been obtained for the Hasvik Gabbro by different workers. The most recent age of 700 ± 33 Ma, based on a Sm-Nd mineral isochron using plagioclase and clinopyroxene, was obtained by combining data from two samples, which have individual age estimates of 706 ± 37 Ma and 688 ± 58 Ma (Daly et al., 1991). By contrast, Rb-Sr mineral dating yielded an age of 528 ± 27 Ma (Sturt, Pringle & Ramsay, 1978). To evaluate these inconsistencies the Hasvik gabbro was selected for U/Pb zircon analysis.

Gabbros RJR-02-30A and 30B were all collected from the Upper Zone of the Hasvik Intrusion, as the more evolved nature of the Upper Zone should have favoured the crystallisation of zircon. These medium grained (± 5 mm) rocks are metamorphosed, with only relict clinopyroxene and orthopyroxene visible in 30A and 30B. Hornblende has replaced most of the pyroxenes in these rocks, along with a small amount of biotite. The samples host a large amount ($\pm 5\%$) of Fe-Ti oxide, both as discrete minerals and as exsolution lamellae in clinopyroxene and amphibole. In 30A and 30B, symplectites of orthopyroxene and magnetite have grown during metamorphism. RJR-02-30D is a metagabbro from the Upper Border series, and comprises plagioclase (55%), clinopyroxene (20%), metamorphic hornblende (20%), with some

quartz and quartz-plagioclase intergrowth. All three rocks proved to be barren of zircon.

Rocks RJR-02-29I and 29J were collected from the Marginal Border Series. These extremely coarse-grained patches are igneous in origin, as they grade into finer-grained gabbro at the edges. These rocks consist of large (3-5 mm) plagioclase crystals (50%), orthopyroxene (20%) and clinopyroxene (25%), with opaque minerals making up the rest of the bulk rock. These two samples are relatively unmetamorphosed compared with the samples from the Upper Zone, and provided zircon for analysis. Unfortunately, the coarse size of the crystals meant that no representative photomicrographs could be taken.

2.1.2. The Lille Kufjord Intrusion, Seiland

The Lille Kufjord Intrusion is found on the island of Seiland (Fig. 2.1, 2.3). This pluton was not dated during this study, but is included here as a documented SIP pluton of supposed tholeiitic composition (Robins & Gardner, 1974). Although less deformed than the Hasvik Intrusion, this layered intrusion nevertheless displays a number of similarities with the Hasvik Intrusion (Robins & Gardner, 1974; Robins et al., 1990), including a degree of contamination by crustal material. The intrusion is about 6.5 km² and is emplaced into the boundary between an older gabbro and gneissic metasediments (Fig 2.3). The intrusion can be subdivided into a Marginal Border Series, marking the edge of the intrusion and up to 100 m thick, and a Layered Series of approximately 1400 m thickness. The Layered Series is divided into a 1130 m thick Lower Zone and a 270 m thick Upper Zone. The Lower Zone is olivine-

bearing, and can be again subdivided into a lower part of a variable cumulate composition, and an upper portion of layered olivine gabbro (Robins et al., 1990). The Upper Zone is composed of layered gabbro, with no olivine present.

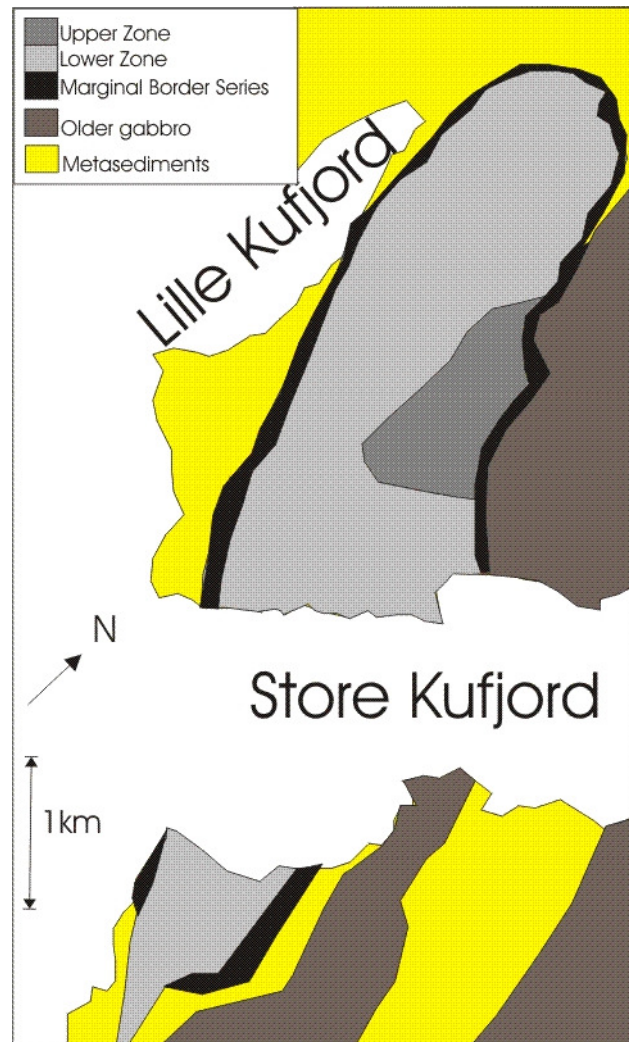


Figure 2.3- The Lille Kufjord Intrusion on Seiland, after Robins & Gardner, 1974.

The Lille Kufjord Intrusion is considered to have been emplaced at mid-crustal depths, under pressures of 5.4- 8.2 kb, based on the presence of kyanite, sillimanite, garnet and cordierite in the aureole (Robins et al., 1990). The only existing age for the intrusion is a Sm/Nd isochron of whole rock, clinopyroxene and plagioclase from the Upper Zone of the intrusion, which yielded an age of 488 ± 57 Ma, which would make this one of the youngest plutons in the SIP (Robins et al., 1990). The layering in

this pluton is considered to be the result of macrorhythmic cycles of convective overturn and magma influx (Robins et al., 1990). In general, most of the work on the pluton has concentrated on the origin of the layering in the cumulate rocks, and the geochemical evolution of the pluton is still poorly constrained (Robins et al., 1990).

2.1.3. The Storelv and Breivikbotn Gabbros on Sørøy

Two other plutons on the island of Sørøy were investigated during this study. The Storelv Gabbro (Fig. 2.1) has been described in two papers (Stumpfl & Sturt, 1965; Sturt & Taylor, 1971). It is a mafic sheet some 40 km long comprising gabbro, along with lesser quantities of diorite and monzogranite, and has long been held to be one of the oldest gabbros (along with the Breivikbotn Gabbro) on Sørøy on the basis of the higher metamorphic grade (amphibolite facies) and degree of folding of the Storelv Gabbro compared with that of the Hasvik Gabbro (Sturt & Taylor, 1971). Both diorite and granodiorite intrude the gabbro, with granodiorite being the last intruded.

Although the mafic rocks are metamorphosed, the extent of deformation in the rocks is extremely variable. In the area where samples RJR-02-37A and RJR-02-37B were taken, the rocks are undeformed. At this location a melanocratic gabbro (RJR-02-37B) is intruded by K-feldspar megacrystic granodiorite (RJR-02-37A). This granodiorite, previously referred to as “adamellite”, can clearly be seen to postdate the gabbro (Fig. 2.4). It forms a network of granitic veins throughout the gabbro, with larger volumes of granitic material intruded at the intersections of several veins. Igneous layering in the gabbro is truncated by the granitic veins. The host gabbro intruded into deformed rocks of the Klubben Psammite, and clearly crosscuts the foliation within the Sørøy

Group. The metagabbro (RJR-02-37B) at this site comprises 50% plagioclase, 20% hornblende, 20% biotite, 5% opaque minerals, 3% clinopyroxene, 1% orthopyroxene, and 1% apatite. The pyroxenes are almost completely metamorphosed to hornblende and biotite, and the plagioclase crystals have been recrystallised. The quartz monzonite (RJR-02-37A) is composed of 50% K-feldspar, 10% quartz, 20% biotite, and 20% hornblende, and zircon can clearly be seen as an accessory phase.

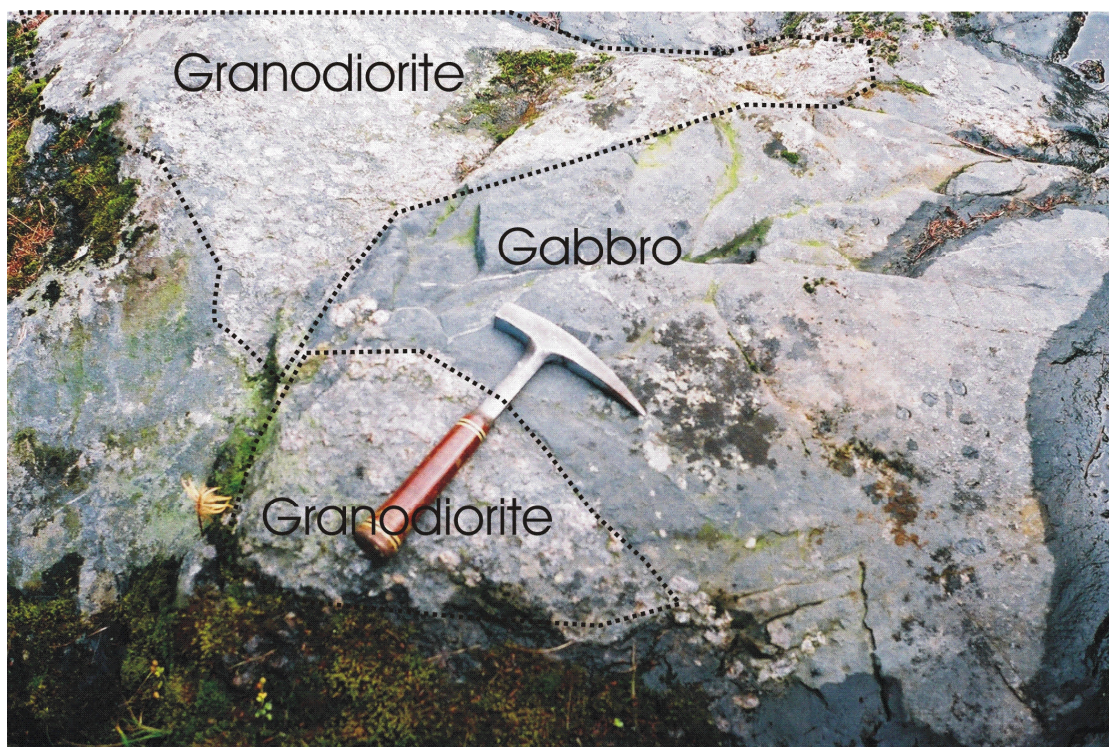


Figure 2.4- Fine-grained Storelv gabbro at Dønnesfjord, Sørøy, intruded by coarse granodioritic magma.

In contrast to the Dønnesfjord locality, 10 km away at Storelv the gabbro and the other intrusions it hosts are highly deformed. Sturt & Taylor (1971) published photographs of foliated gabbro containing boudinaged lenses of monzogranite, along with several other deformation features. It was on the basis of these deformation features that the gabbro was considered syn-tectonic and believed to have been emplaced in the early stages of D2. However, some of the observations made by Sturt

& Taylor (1971) must be questioned. On consideration of the large-scale cross-sections provided by Sturt & Taylor (1971), there does not appear to be folding of the magnitude of D2 within the gabbro (as described by Ramsay et al., 1985). The cross-sections show only small-scale gentle folds of the sort that is normally associated with D3 (Ramsay et al., 1985). Hence, the contention that the gabbro was deformed syn-kinematically with D2 folding is considered suspect and the structural relationships within the Storelv Gabbro should be reassessed in the light of the age dating reported below.

The Breivikbotn Gabbro (Fig. 2.1) is the least characterised of the mafic plutons on Sørøy. It is mentioned in passing in one paper dealing with the alkaline rocks intruded into the pluton and a carbonatite intruded into the nearby country rocks (Sturt & Ramsay, 1965). The pluton is well exposed on the coastal road between Hasvik and Breivikbotn. It is compositionally and texturally heterogeneous, ranging from pyroxenite to gabbro, and from coarse-grained and massive to fine-grained and foliated. Large bodies of diorite are associated with the mafic rocks. Although considered a single intrusion by Sturt & Ramsay (1965), field observations during this study have raised the possibility that the pluton may contain several different intrusions, judging from the variety of mafic rocks present. The sample of metagabbro studied here also displays the growth of orthopyroxene- green spinel symplectites, not found in the tholeiitic gabbros detailed above, but characteristic of the clinopyroxene gabbros detailed below.

Fine-grained mafic dykes are found throughout the pluton, and the density of the dykes is so great in some places that none of the host gabbro is observed. Several

generations of these dolerite dykes are present within the pluton. The situation is further complicated by the presence of large alkaline bodies within the pluton. These rocks range from sedimentary screens and xenoliths altered to alkaline compositions, to nepheline syenite dykes that clearly post-date all other igneous rocks. Significant fenitisation (Sturt & Ramsay, 1965) is associated with these alkaline dykes, altering the mineralogy of both the surrounding gabbro and the country rocks.

The mafic rocks are also metamorphosed. All mafic rocks have been affected by metamorphism of at least amphibolite grade, as indicated by the common occurrence of hornblende within the rocks, and primary features are poorly preserved. In places, large shear zones crosscut the pluton. The pluton has been referred to as a “schistose sheet”, and varies from highly foliated in the north to more massive in the south (Sturt & Ramsay, 1965). Shear zones commonly separate foliated gabbro from massive pyroxenite, and no exposures were found where the original relationships between the different rock types could be evaluated.

RJR-02-31 is an example of amphibolitised gabbro from the Breivikbotn Gabbro. The rock is relatively coarse-grained (± 1 cm) and consists of 40% hornblende, 45% plagioclase, 5% relict clinopyroxene, and 10% symplectic intergrowths. The symplectic intergrowths in this rock differ from those in the Upper Zone of the Hasvik Intrusion, in that these intergrowths comprise green spinel and orthopyroxene, along with some magnetite and serpentine, as opposed to the magnetite-orthopyroxene developed in the Hasvik Intrusion. The presence of serpentine in some of these symplectites is taken to indicate that these symplectites represent a reaction product of olivine. Unfortunately, this rock did not yield any zircon for dating.

Instead of the Breivikbotn gabbro, a tonalite intruded along the eastern edge of the gabbro was investigated. Although poorly exposed, the unit has been mapped by Sturt & Ramsay (1965), who considered it to be genetically related to the mafic pluton, as both bodies have been deformed and metamorphosed by the same deformation events. However, it is possible that the tonalite is significantly younger than the gabbro, so the tonalite provides a lower limit on the age of the Breivikbotn Gabbro.

Outcrop at the sample point (RJR-02-35; Fig. 2.1) was extremely poor. In hand specimen the tonalite is a foliated, medium-grained rock. In thin section, the tonalite comprises plagioclase (50%), quartz (20%), K-feldspar (10%), biotite (15%), and hornblende (5%). The feldspars and quartz are completely recrystallised, as shown by the presence of triple junctions at the edges of the crystals. Biotite and amphibole are present in thick planar bands marking a foliation, and are considered metamorphic in origin.

2.1.4 The Rognsund Intrusion, Seiland

The Rognsund Intrusion straddles the islands of Seiland and Stjernøy (Fig. 2.1), and has been presented as an example of a mafic pluton crystallising from alkali olivine basalt (Robins & Gardner, 1974; Robins, 1982). It is estimated that the intrusion forms a 50 km² oval, much of which is under water (Fig. 2.7). On the Seiland side, the mafic rocks are intruded into metasedimentary gneisses, whereas on the Stjernøy side, the mafic rocks intrude a pre-existing gabbro, the Kvalfjord Intrusion of Daly et al. (1991). This intrusion was not visited during this study.

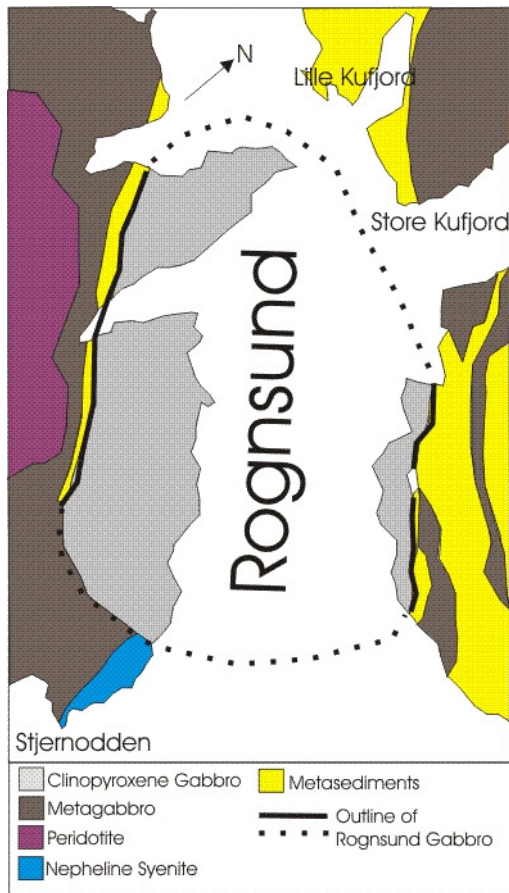


Figure 2.7- Map of the Rognsund Intrusion, straddling the islands of Stjernøy, west, and Seiland, east. After Robins (1982).

The intrusion comprises a Contaminated Zone (<120 m) of banded, olivine-free gabbro containing significant quantities of xenolithic fragments, and a Layered Series of nearly 800 m thickness (Robins & Gardner, 1974; Robins, 1982). The Layered Series is subdivided into a Basal Zone, comprising coarse grained clinopyroxene-olivine gabbros (described as “eucrites” in the literature) and amphibole-plagioclase pegmatites, a Crescumulate Zone, comprising coarse-grained olivine-clinopyroxene gabbro, and a Cumulate Zone, marked by the appearance of magnetite and ilmenite, the increasing abundance of cumulus plagioclase, and a paucity of olivine (Robins, 1982). Plagioclase in the Layered Series is referred to as calcic in composition, and is held to become more sodic in composition with height, changing from An_{85} to An_{65} (Robins, 1982). The Ca-rich clinopyroxenes ($Ca_{47} Mg_{44} Fe_9$ at the base, $Ca_{49} Mg_{35}$

Fe₁₆ at the top) of the Cumulus Zone are depleted in Ti and Al compared to in the lower zones, but, in general, clinopyroxene becomes iron-enriched with increasing stratigraphic height (Robins, 1982). Olivine, however, remains constant in composition, at Fo (mol %) between 74 and 73. The Ca-rich compositions for the clinopyroxenes are considered to be different to the composition of clinopyroxene in tholeiitic gabbros such as those at Skaergaard (Robins, 1982). Also present throughout the intrusion are orthopyroxene-magnetite symplectites, judged to be formed by a subsolidus reaction between olivine and plagioclase (Robins, 1982). The rocks are nepheline-normative throughout the intrusion.

2.1.5. Other gabbros in the SIP

Several other gabbroic intrusions postulated to be of a tholeiitic nature by Robins & Gardner (1974) are exposed in the SIP. One, on the north-eastern edge of Øksfjord (Fig. 2.1), appears only on the geological map of the area (Roberts, 1973). Another, the Hønseby Gabbro on the northern edge of Seiland, is considered to have been partially melted at granulite facies conditions and displays the growth of garnet (Akselsen, 1982). This gabbro may represent a much older episode of magmatism, as it is crosscut by all other SIP intrusions and displays a higher metamorphic grade than the intrusions cutting it (Akselson, 1982). The Husfjord Complex at the southeast corner of Sørøy is another “tholeiitic” gabbro, but has been studied primarily from a structural viewpoint (Speedyman, 1972; 1983). None of these plutons was investigated in this study.

“Clinopyroxene” gabbros are considered to dominate much of the Øksfjord peninsula (Roberts, 1973), although most are not easily accessible from the road (Fig. 2.1). The clinopyroxene gabbros can easily be distinguished from the “syenogabbros” that are found along the eastern side of Øksfjord by the absence of voluminous syenitic/monzonitic sheets interlayered with the gabbro. Several examples of clinopyroxene gabbro were examined, though none provided zircon for analysis.

RJR-02-2 and RJR-02-6A and 6B are taken from the southern end of the Øksfjord (Fig. 2.1). These samples are all gabbroic, and contain predominantly unmetamorphosed clinopyroxene and orthopyroxene, compared to the occurrences listed above. 6A and 6B also contain olivine. 6B is coarse grained (1-2 cm), whereas the other two gabbros are fine- to medium grained. All contain abundant orthopyroxene- green spinel symplectites (Table 2.1.).

RJR-03-113 is taken from a position halfway along Langfjord (Fig. 2.1). Two samples were taken at this location (113A and 113B). Both are metamorphosed at a high grade, but olivine is present in 113B. Both contain abundant orthopyroxene- green spinel symplectites and apatite. These five samples are taken as representative of the mineralogy of the Øksfjord “clinopyroxene” gabbros, although none of these rocks differs in any significant way from other mafic occurrences already detailed.

RJR-03-112 was sampled along the edges of Jokulfjord, in the west of the Øksfjord peninsula. This norite is of a different metamorphic grade from the other samples, and contains poikilitic garnet (5%), along with orthopyroxene (30%), plagioclase (30%), hornblende (25%) and opaque minerals (10%). Although marked on the geological

map for the area as part of the same igneous body as the five samples above, the mineralogy of 112 is significantly different from the other samples. This sample could represent an unrelated intrusion, or simply a higher grade of metamorphism.

Robins & Gardner (1974) describe only one occurrence of their “syenogabbro” type of mafic intrusion, giving an example from Store Kufjord on the island of Seiland. The gabbroic body displayed takes the form of a sheet, in which several layers can be discerned. The Basal Layer (300 m) is gabbroic, contains no olivine and practically no orthopyroxene, and is interlayered with the Klubben Quartzite host. The Middle Layer (500 m) is still gabbroic and olivine-free, but does contain cumulus orthopyroxene. This Middle Layer contains K-feldspar in its upper levels, both as intercumulus and cumulus phases (Robins & Gardner, 1974). The Upper Zone (300 m) is characterised by alternation of syenodioritic, syenitic and perthositic layers containing variable amounts of plagioclase, K-feldspar, orthopyroxene, and clinopyroxene, with apatite, magnetite and sphene as accessory minerals (Robins & Gardner, 1974). These layers, which in some case can be nearly monomineralic, are up to 4 m thick in places. However, little by way of subsequent analytical work has been done on this intrusion, and there is very little information available on “syenogabbros”.

The eastern edge of the Øksfjord peninsula (Fig. 2.1) is formed of mafic rocks of a “syenogabbroic” character- that is, the coast comprises numerous interlayered sheets of gabbro, monzonite, syenite, monzodiorite and other such felsic/syenitic rocks. Little work has been done on differentiating and investigating the different igneous rocks in this area, and the number, size, and stratigraphic variation present in the mafic intrusions are unknown. The relationship between the plutons and the country

rocks is not exposed, except on the generally inaccessible western end of the peninsula. Many of these plutons are layered, and are themselves crosscut by other plutons, hampering the identification of individual bodies (Figure 2.8).

The intrusions are generally considered to be of syn-D2 age (Robins & Gardner, 1974), although little proof has been provided to support this conclusion.

Deformational fabrics within the intrusions are variable, with some rocks showing a foliation, whereas others show no obvious signs of deformation. Under the microscope, most of the mafic rocks display spinel-orthopyroxene symplectites, indicating a reaction involving olivine has occurred after emplacement and crystallisation.

In the first isotopic study of the area, several samples described as “perthosites”, along with gabbro and peridotite, were analysed by Rb-Sr, yielding an age of 612 ± 17 Ma (Brueckner, 1973). At these localities, numerous monzonitic and dioritic bodies are intruded into leucogabbro. Daly et al. (1991) reported a similar Sm/Nd age from mineral separates of 604 ± 44 Ma from a gabbro in the eastern part of the Øksfjord peninsula and an age of 612 ± 33 Ma from a gabbro on Stjernøy. Our sampling along the eastern shore of the fjord was aimed at material similar to that in these previous two studies: RJR-02-3B represents gabbro and RJR-02-40B and RJR-02-41C are monzonitic rocks (Fig. 2.1). RJR-03-101E is an example of a syenitic sheet hosted in the same gabbro.

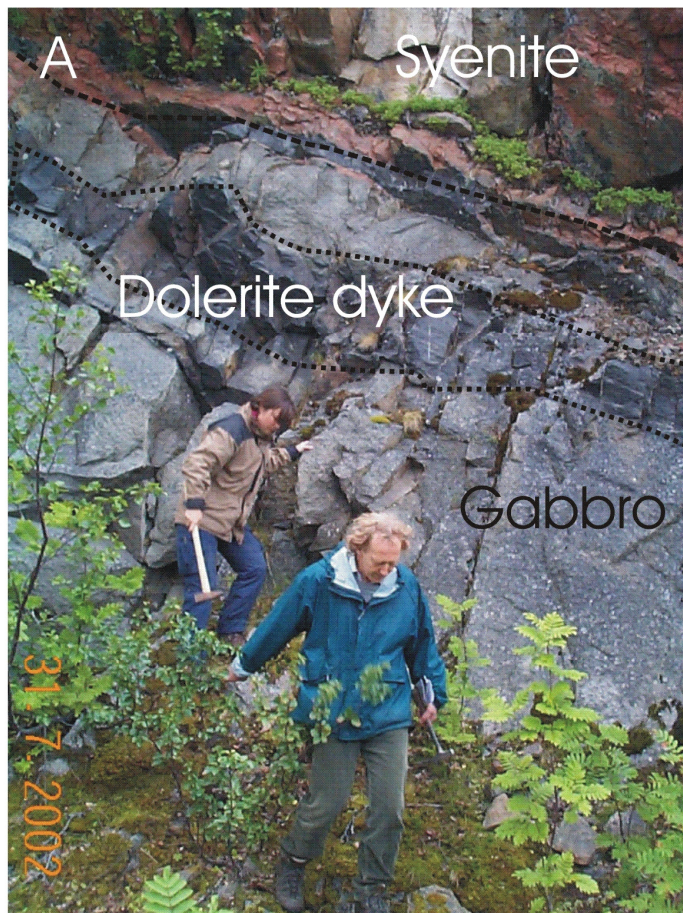


Figure 2.8- A) Syenitic sheet intruded into layered gabbro, Øksjord (site RJR-02-40, Figure 2.1).
 B) Massive gabbro intruded into layered monzonite, Øksjord (site RJR-02-40, Figure 2.1). The gabbro truncates layering in the monzonite.



Figure 2.9- Detail of monzonite- gabbro relationship at site RJR-02-41 (Figure 2.1), Øksfjord. Fingers of monzonite (pale rock) can clearly be seen intruding into the amphibolitised gabbro.

RJR-02-3B is the gabbroic host for the other samples. It is a foliated olivine hornblende metagabbro (plagioclase 20%, clinopyroxene 35%, hornblende 40%, olivine 2%, opaques 2%, orthopyroxene 1%), which contains orthopyroxene-spinel symplectites. The gabbro also contains apatite, pyrrhotite and zircon as accessory minerals. As such, this rock resembles other mafic rocks described above. The samples RJR-02-40B and RJR-02-41C are almost completely composed of feldspar. RJR-02-41C is a monzonite containing perthitic K-feldspar (35%), plagioclase (60%), and quartz (5%), whereas RJR-02-40B is a monzodiorite/monzonite with antiperthitic plagioclase (55%), K-feldspar (40%) and quartz (5%). The monzodiorite also shows intergrowths of quartz and feldspar. Triple junctions formed at the edges of quartz and feldspar crystals in both rocks show that they have been recrystallised, but they are not deformed. Neither rock contains much quartz. RJR-03-101E is an alkali feldspar

syenite and contains 95% K-feldspar, with biotite (2%), clinopyroxene (1%), epidote (1%), and opaques (1%) as minor phases.

Unlike Robins & Gardner (1974), field observations made during this study indicate that some of the syenitic/monzonitic layers in the gabbro are intrusions, rather than cumulus layers in a continuous succession. Furthermore, the veins of such material in the gabbro were considered to be plastically deformed during injection into a hot gabbroic mush (Figure 2.9). Thus, at least some of the monzonitic material was intruded soon after the gabbro, rather than forming in the same magma chamber as the gabbro.

The western coast of Øksfjord is currently undescribed in the literature, and marked as “undifferentiated gabbro” on the geological map of the area (Roberts, 1973). During this study, a number of rocks from this coast were sampled, and several rocks were dated using the U/Pb zircon method. In general, the mafic rocks of this coast are dissimilar to those of the facing coast- there are no significant monzonitic or syenitic sheets present, and the mafic rocks generally comprise homogeneous gabbro. Small quantities of granite and pyroxenite are present.

The samples dated in this study, RJR-03-129C and RJR-03-129D, were taken close to the contact with a large raft of paragneiss (Fig. 2.1), and were selected for the clear structural relationships between the different rocks. Two different mafic intrusions can be seen alongside the road at this location. An orthopyroxenite (RJR-03-129D) clearly intrudes a norite (RJR-03-129C) in this outcrop (Fig. 2.10). A thin chill can be seen at the margin of the orthopyroxenite, and the orthopyroxenite crosscuts the

foliation visible in the norite. At the same place, a small granite body intrudes the mafic rocks. This granite (RJR-03-129A), with an augen gneiss texture, encloses large, rounded xenoliths of mafic material (though not ultramafic material), and is obviously the youngest intrusion in the outcrop (Fig. 2.11.).

Sample RJR-03-129C is a norite, composed of orthopyroxene (40%) and plagioclase (40%), with 15% clinopyroxene, and minor quantities of opaque minerals, and the original igneous assemblage is well preserved. RJR-03-129D is an orthopyroxenite, with 70% orthopyroxene, 20% plagioclase, and minor phlogopite (5%), and does not show the breakdown of orthopyroxene to hornblende commonly seen in other pyroxene- bearing rocks.

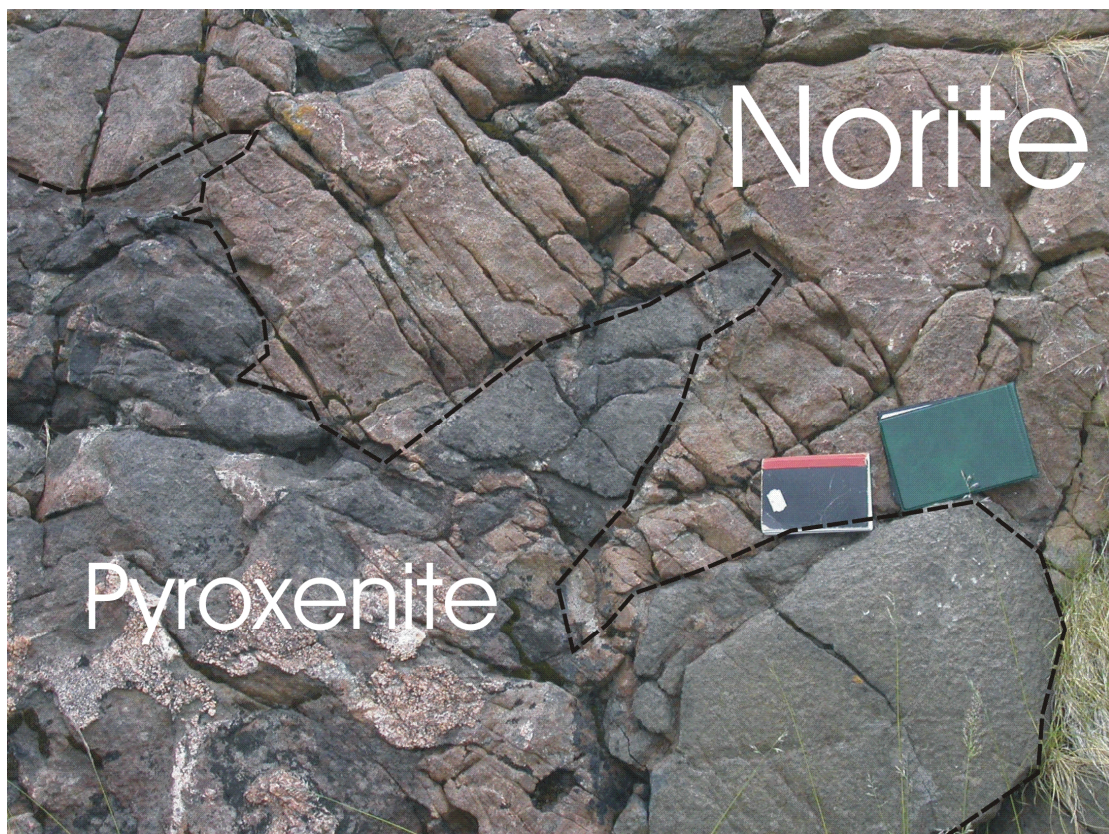


Figure 2.10- Site RJR-03-129, west side of Øksfjord (Figure 2.1). An orthopyroxenite, RJR-03-129D, intrudes a norite, RJR-03-129C.



Figure 2.11- Large (1m) xenoliths of gabbroic rock enclosed within granite gneiss at RJR-03-129.

Numerous other samples were taken along this coast. RJR-03-118, 118B, 119, 120, 125, 130A are all examples of gabbroic and noritic rocks (Table 2.1). There is no olivine in any of these rocks, and amphibolitisation in these rocks is minor compared to other rocks in this study.

2.2. The ultramafic complexes

Bennett et al. (1986) list ten localities in the SIP where ultramafic rocks predominate over rocks of a more gabbroic nature. All of these ultramafic (UM) complexes are emplaced into, and thus post-date, layered mafic intrusions, and are never found intruded into the country rock of the Kalak Nappes. Several of these complexes are greater than 25 km² in area, and form the major bulk of the islands of Stjernøy and Seiland (Fig. 2.1.). The ultramafic rocks are considered to have been emplaced during the D2 phase of Finnmarkian evolution (Bennett et al., 1986). It has been speculated

that the ultramafic rocks represent the fractional crystallisation products of critically undersaturated basaltic magma (Robins & Gardner, 1974; Robins, 1982).

The four major complexes on Seiland, Stjernøy and at Reinfjord are held to show a progressive trend towards more olivine-rich compositions with time (Bennett et al., 1986). Each complex is also closely associated with a pre-existing gabbro, and may have been injected through the same feeder system as the previous magma. Each complex is considered to have been emplaced as a number of successive, mantle-derived magmas emplaced at high temperatures into a crystalline, but not cold, gabbroic pluton, after which processes of fractional crystallisation, assimilation of gabbroic material, and metasomatism contributed to the final composition of the complexes (Bennett et al., 1986).

Ultramafic rocks do not typically contain zircon, and only one small ultramafic complex, at Tappeluft on the Øksfjord peninsula, was dated during the course of this study. However, brief descriptions of the major ultramafic complexes are given below.

2.2.1. UM complexes on Seiland and Stjernøy

The island of Seiland hosts two large ultramafic complexes: the Melkvann UM complex in the centre of the island (Bennett et al., 1986) and the Nordre Bumannsfjord complex in the northern part of the island (Sturt et al., 1980). Both are considered to have been emplaced into their host olivine gabbro intrusions as discordant sheets. The Melkvann UM complex is the largest ultramafic intrusion in the SIP, covering $\pm 100 \text{ km}^2$. The Nordre Bumannsfjord Complex (50 km^2) is very

similar to the Melkvann Complex, and the two complexes may originally have been connected at a higher stratigraphic level. Both complexes comprise numerous crosscutting sheets and dykes of varying composition. Olivine, clinopyroxene, plagioclase, amphibole and spinel are present in varying proportions, though the predominant rock types are olivine clinopyroxenite and peridotite (Bennett et al., 1986). Dunite, wehrlite, and orthopyroxenite can be found as pipes and crosscutting dykes.

The Kvalfjord Complex dominates the eastern side of the centre of Stjernøy (Bennett et al., 1986). This complex is very similar to the complexes on Seiland, comprising predominantly olivine clinopyroxenite and peridotite emplaced in large sheets, with later dunitic and wehrlitic pipes crosscutting pre-existing sheets

2.2.3. Reinfjord UM complex, Jokulfjord

The Reinfjord Complex at the western end of Øksfjord differs from the UM complexes of Seiland and Stjernøy, in that it displays well-developed layering and modal variation in the proportions of olivine, orthopyroxene and clinopyroxene in roughly concentric zones (Bennett 1974; Bennett et al., 1986). The complex is held to comprise two layered series, intruded by a later pluton of dunite and wehrlite. The Lower Layered Series comprises lherzolite, wehrlite and olivine clinopyroxenite, and is separated from the Upper Layered Series by a thin screen of older gabbroic material (Bennett et al., 1986). The Upper Layered Series comprises wehrlite and olivine clinopyroxenite, with well-developed cumulate textures (Bennett et al., 1986).

2.2.4. Tappeluft UM complex, Øksfjord

A small ultramafic complex is found in the southern part of the Øksfjord peninsula, close to the hamlet of Tappeluft (Mørk & Stabel, 1990). This complex comprises coeval pegmatitic gabbro, ultramafic rock and syenite, intruded into a larger mass of clinopyroxene gabbro (RJR-02-6; Fig. 2.12). The pegmatitic gabbro is emplaced as both concordant sheets and discordant dykes and pipes, and comprises large (1-10 cm) crystals of plagioclase, pyroxene and amphibole (Mørk & Stabel, 1990; Fig. 2.13). The ultramafic portion of the complex comprises clinopyroxenites and peridotites, and the associated syenite is coarse-grained (4-5 cm) and composed almost entirely of K-feldspar. All these rocks appear relatively undeformed in hand specimen, but examination of thin sections shows that amphibolitisation and recrystallisation has affected these rocks, although to a lesser extent than in the surrounding gabbros.

RJR-03-109 is a sample of pegmatitic gabbro. This rock is a hornblende gabbro, with large (2-5 cm) plates of igneous hornblende (40%), plagioclase (40%), secondary biotite (10%), and numerous opaques (10%), including ilmenite, magnetite, pyrrhotite and pentlandite. Though appearing undeformed in hand specimen, the plagioclase crystals have been recrystallised and biotite has formed as an alteration around opaque grains.

RJR-02-7 and RJR-03-108A are examples of ultramafic rocks from the complex.

RJR-02-7 is an amphibolite, comprising 60% hornblende, 20% clinopyroxene and 15% plagioclase, with some olivine present (5%) but most converted to orthopyroxene- green spinel symplectites, calcite and serpentine. RJR-03-108A is a harzburgite, with olivine (40%, showing some alteration at the edges), hornblende

(30%), plagioclase (10%) and orthopyroxene (10%). Green spinel (5%) and sulphides (mainly pyrrhotite; 5%) make up the rest of the rock.

RJR-02-8 is the rock that was dated during this study. This rock was retrieved from the area considered as syenite by Mørk and Stabel (1990), and appears as a coarse-grained pink rock in outcrop. This rock is a coarse-grained anorthosite, composed of 90% plagioclase and 10% magnetite-green spinel- orthopyroxene- clinopyroxene symplectites, No nepheline or quartz was identified in the rock.

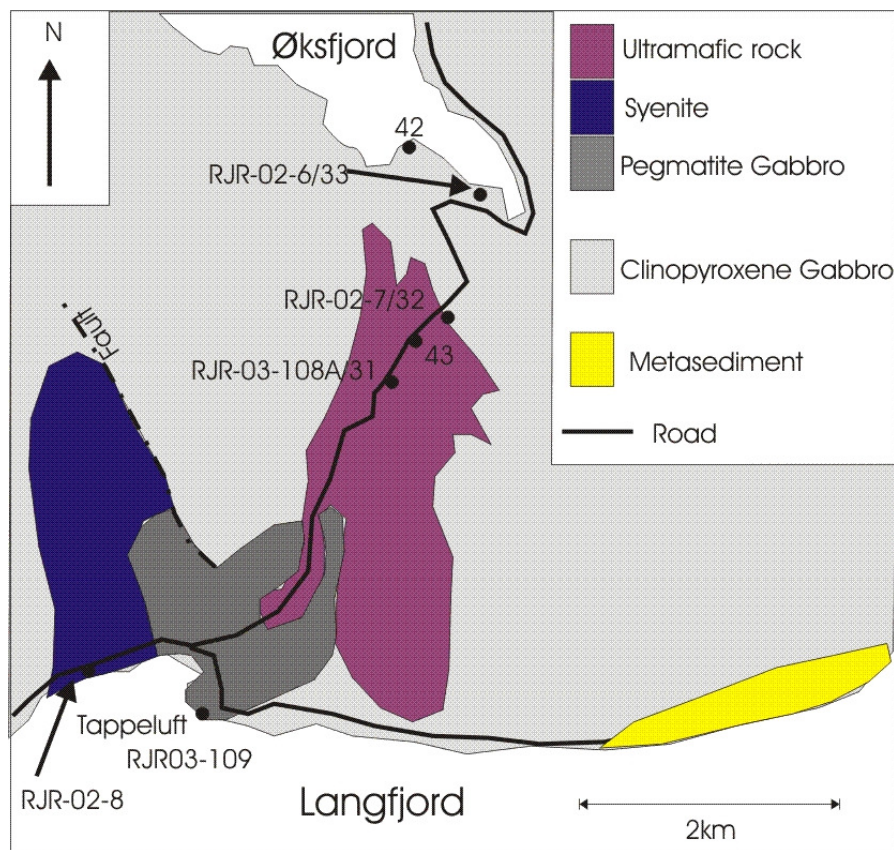


Figure 2.12- The Tappeluft UM complex, after Mørk and Stabel (1990). Sampling points marked with a single number (e.g. 43) are sample numbers from Mørk and Stabel (1990), referred to in Chapter 5. Other sample sites are from this study.



Figure 2.13- Site RJR-03-108, Tappeluft UM complex. Large crystals of primary hornblende with a harrisitic texture can be seen within an outcrop of ultramafic rock.

2.4. Alkaline intrusions

Nepheline syenites and carbonatites are relatively rare rocks, comprising less than 1% of igneous rocks exposed on the surface of the Earth (Sørensen, 1974; Woolley, 1987). However, in the SIP they are relatively common, occurring as large (1-5 km²) complexes and accounting for approximately 5% of the SIP surface exposure. Furthermore, the relatively small (± 5500 km²) area of the SIP hosts four carbonatitic bodies: the Breivikbotn Carbonatite; the Pollen Carbonatite; and the silico-carbonatites of Lillebukt and Store Kufjord (Fig. 2.1). In addition, nepheline syenite dykes are found throughout the exposure of the SIP, including dyke complexes at Lillebukt.

Generally, carbonatites are not suitable targets for zircon extraction, owing to the ability of carbonatitic melt to dissolve large quantities of zirconium without reaching saturation (and thus crystallisation). For the same reasons, nepheline syenites are also commonly barren (Hoskin & Schaltegger, 2003). However, when zircon does occur in these rocks, it is commonly found as large ($> 500 \mu\text{m}$ in diameter) megacrysts (Pedersen et al., 1989; Hoskin & Schaltegger, 2003; Ashwal et al., 2007). Crystals of up to 30 cm are known, and many of the world's gem quality zircons come from alkaline intrusions. The reason for this lies in the relationship between the composition of a melt and its level of Zr saturation. Experiments on a range of synthetic felsic melt compositions revealed a relationship between the temperature of crystallisation, the composition of the melt, and its ability to retain Zr, with composition a more important factor than temperature (Watson & Harrison 1983). Unfortunately the study did not include more extreme melt compositions such as kimberlitic or carbonatitic magmas, but a qualitative extrapolation from the experimental data can be made (Hancher & Watson, 2003).

In general, the ratio of alkaline elements (K and Na) to Al in a melt is the most important factor in determining the ability of a melt to dissolve zircon. Watson (1979) showed that a granitic melt with a $(\text{Na}_2\text{O} + \text{K}_2\text{O})/\text{Al}_2\text{O}_3$ ratio of 2.0 is capable of dissolving up to 3.9 weight % Zr without crystallisation of zircon occurring. The effects of Li, Be, and CO_2 on this relationship are currently unquantified. However, it seems reasonable that the zircon saturation limit of alkaline melts is extremely high, and that zircon precipitation, if it occurs at all, will occur late in the crystallisation of such rocks. Furthermore, any ascending alkaline melt will most likely be undersaturated with respect to Zr, and will dissolve entrained zircons almost

instantaneously (Hanchar & Watson, 2003). One implication of this relationship has been confirmed by textural studies of large zircons *in situ*; large zircons grow extremely fast, and may incorporate other crystals as inclusions (Hoskin & Schaltegger, 2003).

One complication with many alkaline intrusions is that they are associated with significant metasomatism and fenitisation of their host rock (Sturt & Ramsay, 1965; Robins & Tysseland, 1983). Both processes can strongly affect the composition of both the host and the intrusive. For example, the intrusion of carbonatitic material into gabbroic and ultramafic hosts has led to the transfer of many elements between the two rocks, including the addition of Zr to the host rock (Robins & Tysseland, 1983).

Despite these problems, the alkaline rocks of the SIP are an attractive target for zircon extraction, as one study (Pedersen et al., 1989) has already reported U/Pb zircon ages from the Lillebukt Nepheline Syenite Complex on Stjernøy and from carbonatitic dykes on Seiland. In both cases, the zircons were reported as visible in hand specimen, from millimetre scale up to 10s of centimetres in size.

2.4.1. The Breivikbotn Carbonatite, Sørøy

The island of Sørøy hosts an alkaline complex on the edges of a bay on its west coast, south of Breivikbotn (Figure 2.14). This complex comprises a carbonatite intrusion in close proximity to a gabbroic pluton, which is itself intruded by numerous intrusions of syenitic magma (Sturt & Ramsay 1965). Contact between the two distinct alkaline complexes is not observed. However, the spatial proximity of the two intrusions lends itself to the idea that they may be genetically related (Sturt & Ramsay, 1965).

The Breivikbotn carbonatite, and associated shonkinites and syenites, is a deformed narrow (500 m wide) layered intrusive body hosted entirely within the sediments of the Klubben Psammite (Figure 2.14), with best exposure found on the coast. At the northern side of the bay at Haraldseng, the carbonatite has intruded one limb of a north-south fold, with the fold hinge exposed in the cliff face to the west (Figure 2.15). Individual layers of the intrusion run north-south, with the eastern edge marked by a thin (<10 m) aegerine-augite pyroxenite, which appears to lie conformably on the steeply easterly dipping quartzites. Some shearing has taken place along this contact, but there is nothing to contradict the conclusion that this is the bottom contact of the intrusion (Sturt & Ramsay, 1965). The pyroxenite, like the rest of the complex, is intruded by numerous cm-thick nepheline syenite and dolerite dykes, and is extremely variable in appearance.

Overlying the pyroxenite is a coarse grained shonkinite, dominated by feldspars, but also containing pyroxene (aegerine-augite) and amphibole. Also present are calcite, melanitic garnet (Sturt & Ramsay, 1965), titanite and zircon, all of which are visible in hand specimen (<1 mm). The shonkinite has a banded appearance, caused by changes in grain size and mineralogy. The shonkinite is separated from the underlying pyroxenite by a thin band of carbonatitic breccia. This breccia comprises a network of thin carbonatite veins enclosing numerous large angular fragments of both shonkinite and pyroxenite, and is commonly sheared along the edges.

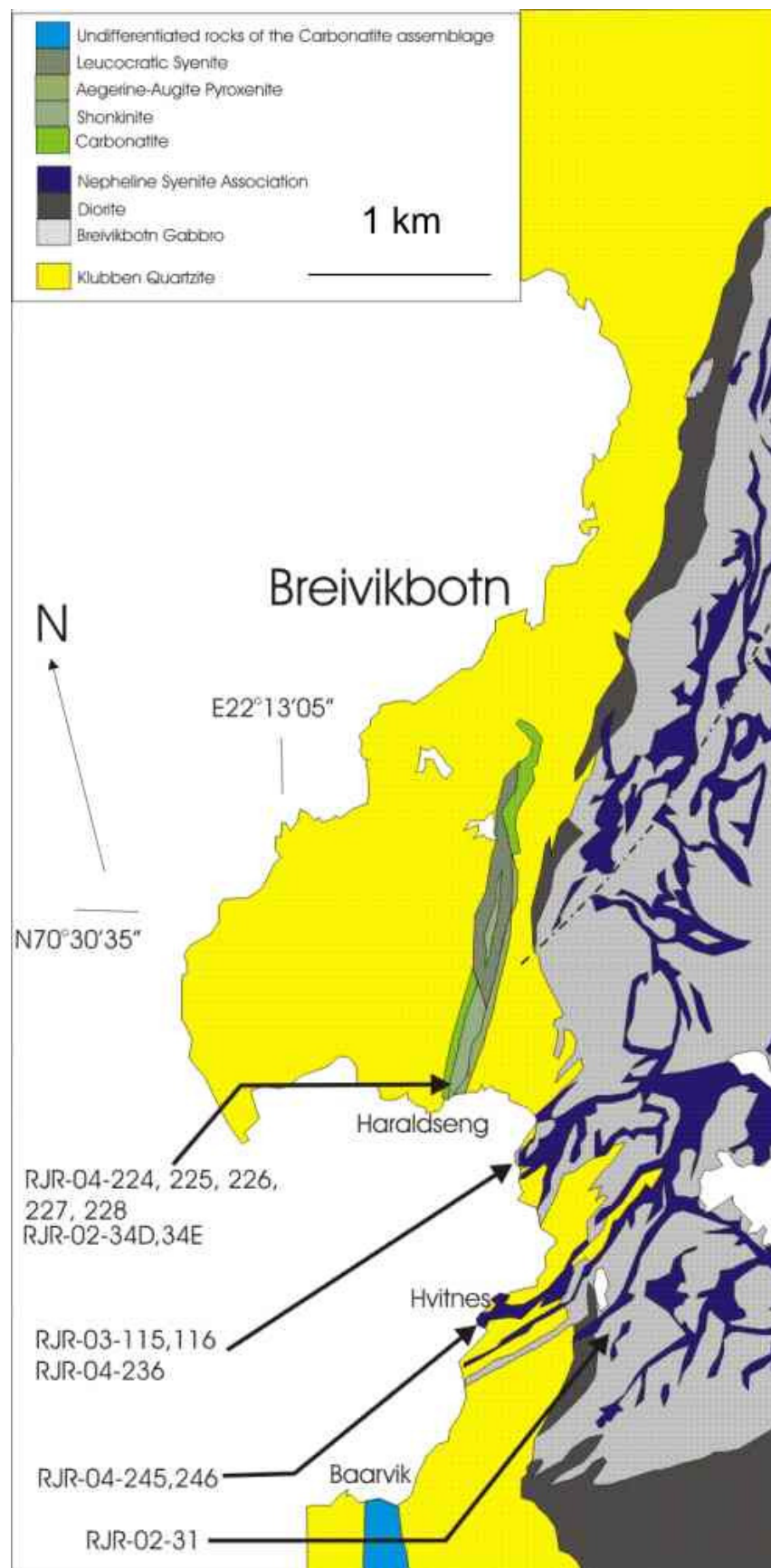


Figure 2.14- Map of the Breivikbotn area, after Sturt & Ramsay, 1965. Sampling points from this study are marked.

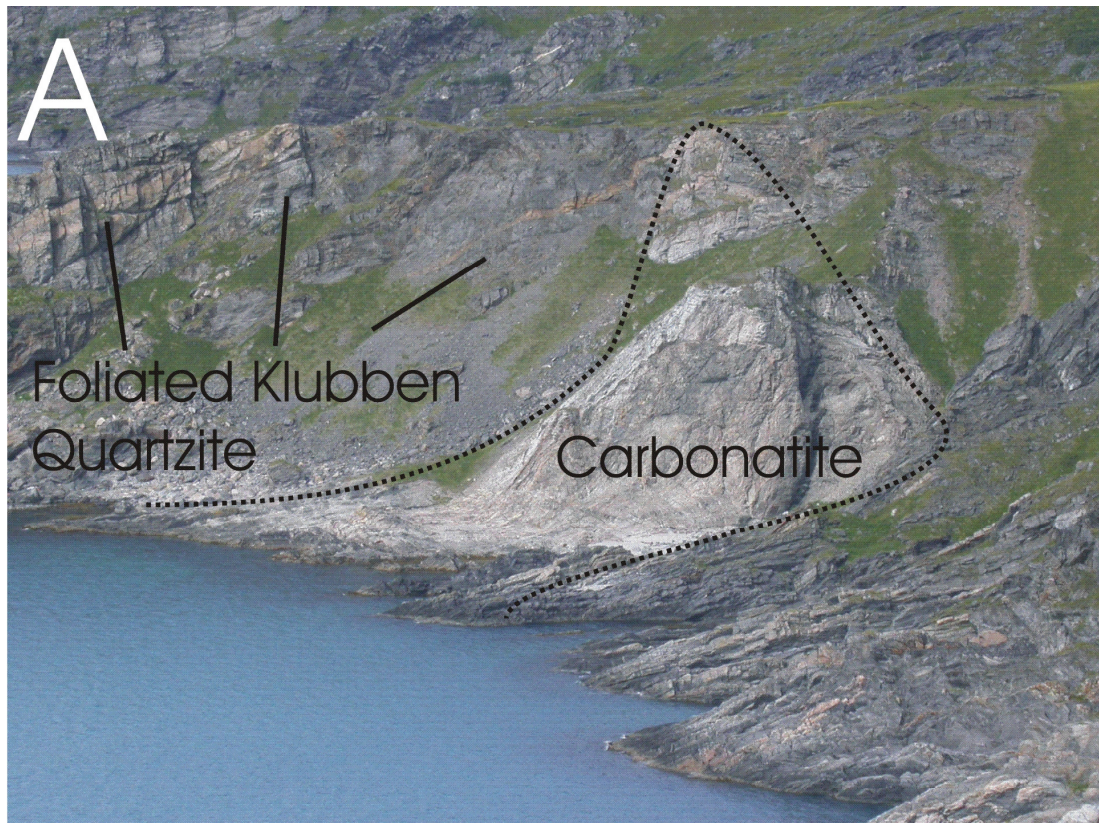


Figure 2.15- A) The Breivikbotn Carbonatite, taken from site RJR-03-116. B) A deformed dolerite dyke within the carbonatite.

The shonkinite grades into carbonatite towards the top of the intrusion, with no clear boundary between the two rock types. The carbonatite is very variable in both texture and composition, and generally occurs as sheets. The rock is composed mainly of carbonate, but there is a significant amount of aegerine, apatite, amphibole and titanite. In some layers, and particularly at the top of the intrusion, the carbonatite contains many fragments of country rock, and is referred to as xenolithic carbonatite (Sturt & Ramsay 1965). There is a sharp contact with the country rock at the top of the intrusion, where the carbonatite truncates a metamorphic foliation in the Klubben Psammities.

An aureole of metasomatic alteration is also associated with the intrusion, especially to the west (top) of the intrusion (Sturt & Ramsay, 1965). The carbonatite is intruded by numerous dolerite dykes (Figure 2.15), which have been used to explain the deformation history of the locality (Sturt & Ramsay 1965). Whereas some dykes are undeformed, most are folded and boudinaged. It was postulated that the dykes intruded the carbonatite prior to deformation, as all folds display the same axis of deformation. On this basis, Sturt & Ramsay (1965) assigned the intrusion of the Breivikbotn carbonatite to the “post-D2” period, and its subsequent deformation to the D3 deformation event.

Numerous samples were taken from the carbonatite, and are listed on Table 2.1.

However, most samples were investigated by Dr. C.Hetherington and photomicrographs are not included for most of the samples. RJR-04-224 is taken from the top of the complex, and comprises a mineral identified as calcite (20%; could be dolomite), scapolite (65-70%), apatite (5%), orthopyroxene (<2%), clinopyroxene

(3%), amphibole (2%) and opaques (<2%). The main carbonate phase, calcite/dolomite occurs as multi-grain patches, with straight grain boundaries straight and perfect triple-junctions. The mineral grains are euhedral, equigranular (~1 mm), with occasional 2-phase fluid inclusions. External grain boundaries are embayed, with concave faces and calcite was replaced by scapolite. Primary orthopyroxene (up to 2 mm in size) is mantled by clinopyroxene. Occasional grains of clinopyroxene are partially replaced by pale brown/yellow amphibole with 120/60 cleavage and strong pleochroism. The scapolite replaces calcite around its edges.

RJR-04- 225 and 227 are garnet-bearing carbonatitic rocks, with two phases of garnet grown. The rocks comprise calcite (45%), scapolite (30%), clinopyroxene (5%), orthopyroxene (7.5%), garnet I (2.5%), garnet II (5%), and opaques (5%). Accessories include amphibole (1-2%), apatite and monazite. Calcite occurs as small patches, occasionally interlinked with one another. Grains can be up to 1.5 mm in size, but are typically around 0.5 mm. Triple junctions and straight edges between calcite grains indicate a degree of recrystallisation, but outer edges of calcite grains in contact with scapolite display concave embayments.

Garnet I has small dark brown cores, anhedral shape, and appears cracked. It contains no inclusions, and is normally ~0.5 mm in diameter, but may range up to 2 mm in diameter. Garnet II, in its most spectacular occurrence, appears as mantle overgrowths to garnet I. This secondary garnet is light brown in colour, possibly andraditic in composition, and displays no cracks or fractures, but hosts multiple inclusions, including calcite and occasionally opaques. Some of the opaque mineral grains are also seen to be mantled by an apparent overgrowth of garnet. Garnet II is the more

voluminous of the two types of garnet, and occurs as discrete crystals too.

RJR-226 is an example of a shonkinite. This is a K-feldspar- dominated rock (70%), with equidimensional grains of K-feldspar (0.25 mm) commonly displaying rounded crystal edges, suggesting some recrystallisation. Occasional large phenocrysts (4.5-5.0 mm) of plagioclase display well developed perthite textures. This rock has a low carbonate content (5%), the grains of which are mainly found in the matrix of plagioclase in skeletal form, suggesting they grew as intercumulus phases. Some carbonate can also be found as larger (0.5-0.7 mm) subhedral crystals associated with scapolite (25%). This rock also hosts angular fragments of zircon, which sometimes show euhedral crystal faces. These zircons are generally heavily cracked and fractured.

RJR-228 is another carbonatitic rock, and is extremely friable in hand specimen. The rock comprises carbonate (>60%), orthopyroxene (20%), clinopyroxene (5%), opaques (10%), and apatite (1-2%), but does not contain any felsic minerals or scapolite. Carbonate minerals display the growth of triple points, euhedral undeformed grains, and straight grain boundaries. The mafic minerals are dominated by orthopyroxene (light green, faint pleochroic), with limited overgrowth of dark green clinopyroxene with strong pleochroism. This is probably the only section that is a true carbonatite (>50% calcite by mode), but interestingly displays the lowest grade of metamorphism.

2.4.2. The Breivikbotn Syenitic Complex, Sørøy

The syenitic magma complex in the Breivikbotn area is much more extensive and involves a much greater volume of alkaline magma than the Breivikbotn Carbonatite. This alkaline magma intrudes the Breivikbotn Gabbro and its surrounding country rock (Sturt & Ramsay, 1965), and consists of a stockwork of veins and sheets (Sturt & Ramsay 1965).

Sturt & Ramsay (1965) described the alkaline magmatism as a single magmatic event. However, the alkaline magmatism consists of numerous discrete intrusions of differing composition, which are themselves intruded by nepheline syenite dykes, indicating a multi-stage emplacement history. Compositions range from nepheline-bearing to nepheline-free syenites, with both coarse- and fine-grained varieties observed. K-feldspar- magnetite pegmatites are a third coeval but rarer type of intrusion in the area. Also associated with the alkaline rocks are numerous metasomatically altered fragments of country rock and gabbro, which can be difficult to separate from the igneous rocks without petrographic examination of thin sections of the rock in question (Sturt & Ramsay, 1965).

The oldest alkaline dykes are highly deformed, have a gneissic texture, and sometimes show intense folding (Figure 2.16). A blue colour depicts a foliation expressed by biotite, with some amphibole, in a generally white matrix. In shear zones, sodalite may be developed, and cancrinite may replace nepheline. Younger dykes cutting the older deformed gneissic alkaline rocks and host gabbros have straight, undeformed edges and sharp contacts.

Three localities, representative of the general occurrence of alkaline dykes, were selected for sampling. Above Haraldseng (Figure 2.14), two examples of alkaline gneiss were collected (RJR-03-115 and RJR-03-116). Both these rocks have been intruded into the Breivikbotn Gabbro. RJR-03-115 is composed of 40% K-feldspar, 20% nepheline, 20% plagioclase, 10% muscovite, 5% biotite and minor opaques (3%) and epidote (2%). Bands of recrystallised felsic material form a foliation in the rock. RJR-03-116 is similar, but contains more nepheline (30%) and less K-feldspar (30%).

Close to this location, a nepheline syenite dyke, crosscutting both alkaline gneiss similar to RJR-03-116 and the host gabbro, was sampled for dating (RJR-04-236; Fig. 2.14). This dyke is 40 cm wide and shows an alignment of biotite crystals parallel to the walls of the dyke. The dyke coarsens towards the middle, with biotite being concentrated along the sides. Modally, this rock is identical to RJR-03-115.

At Hvitnes, numerous alkaline dykes intrude the Klubben Psammite. Many pictures of this locality can be found in Sturt & Ramsay (1965). At one location, deformed alkaline gneiss (RJR-04-245) and country rock are crosscut by a thin (0.4 m), apparently undeformed dyke of biotite-free syenite (RJR-04-246). RJR-04-245 comprises K-feldspar (90%), nepheline (4%), hornblende and orthopyroxene (1%), opaques (4%), muscovite (1%), plus accessory phases such as zircon and apatite.

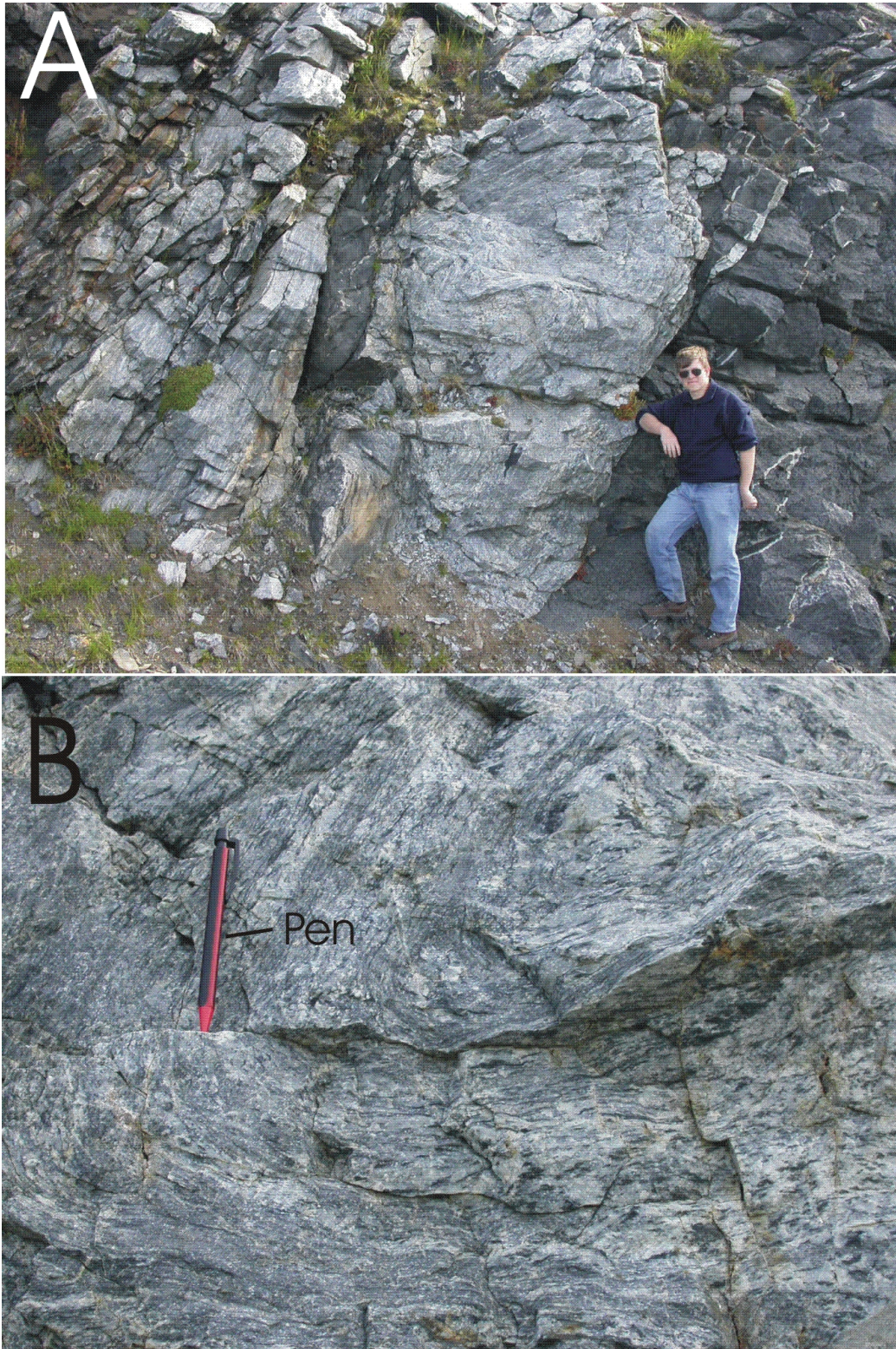


Figure 2.16- A) Alkaline gneiss intruded into the Breivikbotn Gabbro. B) Close-up of alkaline gneiss, showing the deformed bands of biotite that appear bluish in hand specimen.

The feldspar- nepheline matrix is generally equigranular (0.5 mm), with abundant triple junctions. Some areas of feldspar have a fuzzy-yellow appearance caused by sericitisation, indicating the passage of fluids through the rock. Mafic minerals are minor. Hornblende is definitely present (perfect cleavage in two grains), showing yellow to green pleochroism. Another grain is dark green and displays no cleavage, and has been identified as clinopyroxene. All grains are euhedral and most likely magmatic in origin.

The nepheline syenite dyke intruded into this rock, RJR-246, comprises plagioclase (25%), nepheline (25%), orthoclase (40%), biotite (8%) and muscovite (2%), with accessory zircon clearly visible. A weak foliation is delineated by biotite, and to a lesser extent by muscovite. Furthermore, many of the plagioclase crystals in the rock are aligned with their twin planes parallel to the foliation marked by the micas, while other grains of plagioclase are aligned perpendicular to the foliation. Few perfect triple junctions are present and there is only limited sericitisation of the feldspars.

2.4.3. Store Kufjorden, Seiland

Seiland hosts several large gabbro plutons (Robins, 1982; Robins et al., 1990), into which two large ultramafic bodies were intruded (Sturt et al., 1980; Bennett et al., 1986). The intrusions are generally held to be emplaced into Eidvågeid Gneiss (Akselson, 1982). An occurrence of large, occasionally gem quality, zircons in a nepheline syenite dyke at Store Kufjord was dated by U/Pb ID-TIMS to 531 ± 2 Ma (Pedersen et al., 1989), and it was deemed necessary to visit and compare the occurrence of zircon at this location with that at Breivikbotn.

In Store Kufjord (Figure 2.1) several large nepheline syenite dykes cut the mafic and ultramafic plutons. The dykes form an *en echelon* array of near vertically-dipping sheets, orientated east-west. They range in thickness from 30 cm to 10- 15 m, and the larger intrusions can be traced along strike for several hundred metres. In this area, the host rocks are a gabbro-norite, and an ultramafic complex. The ultramafic complex, referred to as the Melkvann UM, intrudes the gabbro-norite, and comprises olivine-pyroxenite and peridotite, with isolated occurrences of dunite and wehrlite (Bennett et al., 1986). The nepheline syenite dykes are the youngest igneous phase in the area, crosscutting the well developed rhythmic layering in the ultramafic rocks (Bennett et al., 1986).

The nepheline syenite dykes are very coarse grained and comprise predominantly nepheline and feldspar crystals up to 10 cm in length. Biotite is an important component of most dykes, although the concentration varies greatly along strike. The biotite can be found as thick books, up to 5 cm thick and often 10 cm or more in length. Pedersen et al. (1989) suggested that the dykes were deformed, but, whereas there is occasionally a tectonic foliation and some mylonitisation near the edge of the dykes, no conclusive evidence for internal deformation and/or metamorphism was observed in this study.

Zircon megacrysts (Pedersen et al. 1989) are found in amphibolitised Melkvann peridotites at the edge of a large nepheline syenite dyke and occasionally in the nepheline syenite where fingers of amphibolitised peridotite extend into the nepheline syenite. The zircons are brown euhedral crystals and up to 5 cm in length. The occurrence of the zircons in the ultramafic rock surrounding the nepheline syenite

dykes is significant. Textures indicate that they probably formed by a reaction between the alkaline intrusions and their host rock, possibly similar to the process reported from the Pollen carbonatite on Stjernøy (Robins & Tysseland, 1983), rather than crystallising in the nepheline syenite itself. Zirconium has thus been transported from the undersaturated nepheline syenite magma and released into the surrounding ultramafic rock. The magmatic fluids have extensively altered the peridotite to amphibole, releasing SiO₂ which has been scavenged by the Zr for the growth of zircon. The metasomatic reaction would have been facilitated by high temperatures, and must have occurred contemporaneously with the intrusion of the nepheline syenites and associated alteration of the ultramafic rock. Thus, the age reported by Pedersen et al. (1989) still represents the age of intrusion of the nepheline syenites.

2.4.4. Lillebukt, Stjernøy

Pedersen et al. (1989) also presented a U/Pb age of 523 ± 2 Ma for zircon, supported by coeval but low-U titanite, for the Lillebukt Alkaline Complex on the west coast of Stjernøy (Fig. 2.1). Originally described by Heier (1961, 1964), with further contributions from Robins, (1980), Skogen (1980), and Cadow (1993), the Lillebukt Alkaline Complex is the largest body of alkaline rock in the SIP. It comprises approximately 13 km² of nepheline syenite and calcic syenite, emplaced in a roughly concentric fashion. The host rocks comprise gabbro in the north and east and amphibolitised pyroxenite (hornblendite) to the west of the complex.

The southern portion of the complex is composed of a mass of nepheline syenite sheets and lenses, where a nepheline mining operation is active (Heier, 1964). The

northern portion of the complex consists of a calcitic syenite, generally described as carbonatite (e.g. Cadow, 1993) but lacking the 50% carbonate content by mode required to be a true carbonatite. This calcitic syenite intrudes the nepheline syenites. The older intrusions of the alkaline complex are dolerite dykes (Cadow, 1993), that have in turn been truncated by later alkaline intrusions, including a mass of apatite-rich hornblende clinopyroxenite dykes. The whole complex is reported to have been deformed inhomogeneously, as portions of the complex display folded compositional banding and a mineral foliation (Skogen, 1980). However, the host hornblendite is relatively undeformed, indicating that deformation strain has been accommodated in the more ductile alkaline rocks.

The zircons reported by Pedersen et al. (1989) came from nepheline syenite dykes cutting locally deformed nepheline syenite and metasomatized gabbro, but no specific description of the locality is provided. The zircons gave a concordant U/Pb TIMS age of 523 ± 2 Ma. During this study, numerous small elongate brown zircons of up to 1 cm in length were found in hornblendite adjacent to nepheline syenite in the northwestern part of the complex, in a manner similar to that observed at Store Kufjord on Seiland. Without a detailed field and sample description, the significance of the existing U/Pb ages is limited and we are restricted to concluding that the Lillebukt Alkaline Complex was emplaced in stages, one of which has been dated to around 523 Ma.

2.4.5. Other occurrences of alkaline rocks in the SIP

The carbonatitic occurrence at Pollen on Stjernøy could not be visited during the course of this study. However, field descriptions of the locality imply that the carbonatite is similar to that at Breivikbotn (Robins & Tysseland, 1983). Other nepheline syenite dykes occur throughout the complex, and many were examined. However, these dykes are texturally similar to those on Sørøy or Seiland. No dykes of obviously older ages were found.

Chapter 3: Analytical Methodology

3.1. U/Pb ID- TIMS analysis

All U/Pb analyses in this study were conducted by isotope dilution thermal ionisation mass spectrometry (ID-TIMS) at the University of Oslo, following the methods established by Krogh (1973). The samples were crushed using a Retsch swing mill, which was rigourously cleaned after each sample, and the finer particles were then removed from the powder using a Wifley shaker table. Zircons and monazites were extracted from the powders by a two-stage heavy liquid process (Krogh, 1973). Zircons were hand picked on the basis of morphological observations. After hand picking, the selected zircons were abraded in an air-abrader to remove altered rims (Krogh, 1982). Both zircons and monazites were then cleaned thoroughly with dilute nitric acid, followed by deionised water and acetone rinses. The zircons were then dissolved in Teflon bombs and the monazites in Savillex vials, along with a mixed spike of ^{205}Pb - ^{235}U . After dissolution, the solutions were passed through ion exchange columns in order to remove Zr, Hf, and the rare earth elements, which can potentially interfere with the analytical signal obtained from the sample (Corfu & Noble, 1992). The samples were loaded on zone-refined Re filaments with Si-gel and H_3PO_4 and measured on a MAT 262 mass spectrometer either on Faraday cups in static mode, or, for smaller samples and all $^{207}\text{Pb}/^{204}\text{Pb}$ ratios, by peak-jumping in an ion-counting secondary electron multiplier. The secondary electron multiplier data were corrected for a non-linear bias using an exponential equation whose parameters were adjusted based on concurrent measurements of the NBS982 Pb standard. In addition, all the data were corrected for 0.1%/a.m.u. fractionation using reproducibility factors of 0.05%/a.m.u. for Faraday data and 0.1% /a.m.u. for secondary electron multiplier

data. Corrections for isotopic fractionation and SEM analytical bias were performed prior to correcting the analyses for the analytical blank (2 pg Pb and 0.1 pg U) and initial common Pb content (Stacey & Kramers, 1975).

Final analysis and presentation of the data was conducted using the program ISOPLOT (Ludwig, 2003). In many cases, the array of concordant zircon analyses requires a synthetic anchoring point to be added in order to calculate an age. It may be that an anchoring point at 420 ± 20 Ma, the Scandian age of Caledonian orogenesis (Gayer, Hayes & Rice, 1985), would be a reasonable age for the alteration and growth of metamorphic overprints on the zircons, but this is currently an assumption. The use of an anchoring point of 200 ± 200 Ma covers a multitude of possibilities, eliminating bias on the part of the analyst. All errors are reported at the 2-sigma confidence interval, except where noted.

3.2. MC-ICP-MS analysis for zircon Hf isotopic signatures and whole rock Sm/Nd

The rare-earth separates obtained from zircon during anion exchange chemistry on zircons for ID-TIMS were later analysed for their $^{176}\text{Hf}/^{177}\text{Hf}$ content at the University of Bern, Switzerland, under the supervision of Prof. Jan Kramers. The separates were taken up in 1 N HCl -0.1 N Hf acid and run through a single-stage anion exchange column (Column A of Patchett and Tatsumoto, 1980) to separate out Hf from the bulk of the REE elements. Hf is a stoichiometric element in zircon, and can comprise between 0.5 and 5% by mass of the zircon. Comparatively, the 20-200 ppm of Lu present in zircon is insignificant, and little or no back calculation for the decay of ^{176}Lu to ^{176}Hf is required to use the $^{176}\text{Hf}/^{177}\text{Hf}$ ratio. Owing to the relatively recent age of the zircons in this study, Lu was not measured. The resultant Hf solute was

analysed on a Nu Plasma multi-collector ICP-SFMS, with care taken to measure potential interferences on the Hf signal (Goolaerts et al., 2004). No measurable interferences were detected, and the $^{176}\text{Hf}/^{177}\text{Hf}$ values are presented in Chapter 5 with no corrections. The zircon standard 91500 was run five times during the analytical run, and the $^{176}\text{Hf}/^{177}\text{Hf}$ content reported by the instrument was within error of the reference values for the standard reported by Goolaerts et al. (2004).

Eight rocks were also prepared and analysed at the University of Bern for their Sm/Nd isotopic content. Daniel Rufer assisted in the sintering and chemical preparation of the samples according to the method presented in Kleinhanns et al. (2002). The separates were then analysed on a Nu Plasma MC-ICP-SFMS, with care taken to correct for potential interferences from the rare earths. An inhouse standard was run with each batch to correct for analytical drift. The raw data from the runs was corrected for interferences from Ce and Sm (Nd), and for Gd (Sm), and a Q factor was also applied (Kleinhanns et al., 2002).

3.3. Major and trace element analysis

A set of samples was analysed for major and trace element geochemistry at the University of KwaZulu-Natal, under the supervision of Prof. Allan Wilson. Major element content was determined by ED-XRF on fused glass discs, and trace elements were analysed on an ICP-MS. The XRF analyses were measured against an inhouse standard by the laboratory, and a FeO/Fe₂O₃ ratio of 0.1 was assumed for the purposes of CIPW norm calculation.

These samples were originally digested by acid microwave dissolution, but the Zr content returned was not in accord with the observed zircon content. Since it is known that acid attack (even high temperature and high pressure) dissolutions may not dissolve highly refractory phases (such as zircon), the samples were re-digested using several different types of fluxes. Most fluxes were largely found to have unacceptably high blank levels. Others caused high solute contents. The samples were analysed five times with these different fluxes. Apart from the blank levels, a problem with flux dissolutions is that the added solute content from the flux results in very small amounts of sample effectively being analysed (usually about 12 mg). One flux was finally found to produce clear solutions with complete sample dissolution. Concern was expressed that using HF together with the flux produced insoluble fluorides of some elements which caused co-precipitation of some elements (including the REE), so dissolution of the fluxed sample was carried out using only nitric acid. However, this was observed to cause precipitation of some elements in solution (Nb, Ta and W), by comparing with international standard materials. Base metals and selected other metals (Ni, Cu, V, Cr, Sc and P) appear to have been contaminated from the memory effect of the Pt-crucibles used for the fusions, since such metals (from previous digestions of chromite ore) are ingested into platinum crucibles and may therefore contaminate subsequent samples. Co does not appear to be affected, whereas Cr and Cu are by far the worst elements in this regard.

Fusions produced greatly increased Zr contents for some samples confirming that refractory mineral components (such as zircon) were not dissolved in the microwave digestion. Others were very similar to the original (microwave dissolutions) determinations. The extreme range of some elements (Zr and Sr) requires samples to

be re-run at different dilutions, and these results were confirmed with XRF analyses of pressed powder pellets. The data from three international standards run alongside the ICP-MS samples is presented in Table 3.1.

Table 3.1 International Reference Materials for Trace Element Analyses

	Measured BCR-1	Recommended BCR-1	Measured BHVO-1	Recommended BHVO-1	Measured BIR-1	Recommended BIR-1
Rb	47.200	47.200	9.459	11.000	0.261	0.250
Sr	331.796	330.000	400.830	403.000	108.419	108.000
Y	37.471	38.000	27.998	27.600	16.935	16.000
Zr	189.927	190.000	179.079	179.000	16.299	15.500
Ba	695.673	681.000	136.153	139.000	6.817	7.000
La	25.073	24.900	15.692	15.800	0.811	0.620
Ce	53.605	53.700	39.069	39.000	1.971	1.950
Pr	6.950	6.800	5.580	5.700	0.394	0.380
Nd	28.653	28.800	25.336	25.200	2.579	2.500
Sm	6.669	6.590	6.128	6.200	1.122	1.100
Eu	1.993	1.950	2.016	2.060	0.513	0.540
Gd	6.818	6.680	6.274	6.400	1.685	1.850
Tb	1.056	1.050	0.955	0.960	0.367	0.360
Dy	6.232	6.340	5.292	5.200	2.597	2.500
Ho	1.261	1.260	0.989	0.990	0.589	0.570
Er	3.494	3.630	2.497	2.400	1.680	1.700
Tm	0.530	0.560	0.350	0.330	0.273	0.260
Yb	3.349	3.380	2.039	2.020	1.722	1.650
Lu	0.508	0.510	0.291	0.290	0.266	0.260
Hf	4.865	4.950	4.459	4.380	0.632	0.600
Th	5.519	5.980	1.179	1.080	0.033	0.030
U	1.683	1.750	0.437	0.420	0.017	0.010
Nb	13.667	14.000	19.475	19.000	0.666	0.600
Ta	0.842	0.810	1.185	1.230	0.045	0.040
W	0.400	0.440	0.300	0.270	0.187	0.100
Pb	12.669	13.600	2.807	2.600	3.768	3.000

Chapter 4: U/Pb zircon ages from the Seiland Igneous Province

4.1 Introduction

The need to determine accurate and precise formation ages for rocks or structures, and the rates at which geological processes occur, pervades nearly all aspects of Earth science research. An accurate and reliable geochronological framework is especially required for the interpretation and understanding of regions with complex tectonic and magmatic histories to untangle problematic sequences of events. Wilson-cycle tectonics postulates the opening of oceanic basins through intracontinental rifting, a period of mid-ocean ridge spreading that is later reversed, with final closure of the basin taking place during continental collision along the original margins of the rift (Wilson, 1966). Subsidiary settings, such as island arc tectonics or back-arc extension, are variations on this overarching theme. More modern versions of plate tectonic theory emphasise that plates can rotate and move significant distances, and a rifted margin need not collide again with the margin from which it rifted (e.g. Hartz & Torsvik, 2001). In the global geodynamic framework, orogenic belts may contain fragments of previous tectonics settings, such as the original rift margin. However, recognising these fragments can be difficult in highly metamorphosed and deformed orogenic belts.

The SIP contains numerous intrusions, including numerous alkaline intrusions. In general, voluminous alkaline magmatism is considered to be an indicator of tectonic extension (Bailey, 1974; Bell and Blenkinsop, 1987; Burke et al., 2003), so determining the relationship between the carbonatite and alkaline intrusions and the

SIP gabbros and ultramafic rocks is important for understanding the pre-Caledonian history of this region of Norway.

Geochronology from several mafic, felsic and alkaline bodies are presented here. The primary objective of this study is to determine the chronological relationship between the alkaline and mafic rocks, and place greater constraints on a tectonic framework for the magmatism in the SIP.

4.2. Results of U/Pb age dating

4.2.1. Hasvik Intrusion, Sørøy

The Hasvik Intrusion has been described in Chapter 2. Two rock samples were collected for analysis. RJR-02-29I and 29J were collected from the Marginal Series, along the road at the southwestern edge of the pluton (Figure 2.2). Both samples were collected from small (<0.5 m) pegmatitic patches in the pluton. These coarse-grained patches are igneous in origin and grade into finer-grained gabbro at the edges. These rocks consist of large (3-5 mm) plagioclase crystals (50%), orthopyroxene (25%) and clinopyroxene (20%) , with opaque minerals making up the rest. The pegmatites appear undeformed, and no significant differences exist between the two samples.

The results of the U/Pb analyses are shown in Table 4.1. The two samples (RJR02-29I and 29J) each provided ± 20 zircons, comprising mostly fragments. The zircons are generally yellow-brown, and show good crystal shape. Whole grains are slightly rounded and up to 200 μm wide. The morphology of the crystals is not generally indicative of a metamorphic origin, as the zircons do not show obvious cores under the microscope or under CL, and do not show rounded edges indicating resorption.

Table 4.1: U/Pb analyses of various rocks from the Seiland Igneous Province

Sample	No.	Type	Weight [ug]	U [ppm]	Th/U ^a	Pb com ^b [pg]	²⁰⁶ Pb/ ²⁰⁴ Pb ^c	²⁰⁷ Pb/ ²³⁵ Pb ^d	²⁰⁶ Pb/ ²³⁸ Pb ^d	rho
Hasvik Gabbro RJR02-29J	A	2 clear rounded	1	1424	0.41	7.2	1146	0.74046 ± 0.00470	0.09145 ± 0.00052	0.77
	B	1 brown euhedral	1	447	0.39	0.9	2973	0.75086 ± 0.00424	0.09275 ± 0.00050	0.80
	C	6 clear frags	9	55	0.43	1.7	1691	0.74499 ± 0.00502	0.09228 ± 0.00050	0.69
	D	4 brown frags	3	446	0.49	1.0	7679	0.75015 ± 0.00428	0.09222 ± 0.00051	0.91
	E	1 clear rounded	1	1128	0.95	0.6	10152	0.74813 ± 0.00301	0.09215 ± 0.00036	0.84
	F	4 large brown euhedral	10	213	0.87	4.0	3143	0.74683 ± 0.00369	0.09272 ± 0.00028	0.62
Hasvik Gabbro RJR02-29I	A	4 brown rounded	143	1375	0.80	38.9	28492	0.72980 ± 0.00270	0.08995 ± 0.00032	0.98
	B	1 brown rounded	82	231	0.58	12.1	8941	0.73800 ± 0.00260	0.09084 ± 0.00031	0.67
	C	3 brown rounded	95	779	0.42	8.5	48827	0.72740 ± 0.00250	0.08962 ± 0.00029	0.98
Breivikbotn Diorite RJR02-35	A	13 blocky frag	13	238	0.36	14.4	1261	0.75328 ± 0.00385	0.09259 ± 0.00037	0.79
	B	14 large brown tips	100	307	0.36	465.8	401	0.76029 ± 0.00624	0.09281 ± 0.00033	0.44
	C	Numerous clear tips	15	329	0.37	4.7	6151	0.75209 ± 0.00279	0.09227 ± 0.00030	0.87
	D	Numerous small tips	72	300	0.37	2.6	47707	0.76038 ± 0.00450	0.09282 ± 0.00055	0.93
	E	Numerous clear frags	20	437	0.39	3.9	12843	0.74199 ± 0.00454	0.091073 ± 0.00063	0.85
Storelv Gabbro RJR02-37B	A	1 large brown euhedral	1	6650	0.19	2.0	19385	0.73752 ± 0.00395	0.09072 ± 0.00051	0.92
	B	1 large clear euhedral	1	15624	0.88	24.6	3700	0.75406 ± 0.01501	0.09264 ± 0.00184	1.00
	C	1 large brown euhedral	1	4488	0.99	1.0	25068	0.74742 ± 0.00462	0.09175 ± 0.00056	0.98
	D	1 brown euhedral	1	12979	1.31	2.4	30935	0.73756 ± 0.00278	0.09075 ± 0.00033	0.98
	E	1 brown euhedral	1	8418	0.90	1.9	26054	0.74949 ± 0.00327	0.09187 ± 0.00040	0.96
Storelv Granodiorite RJR02-37A	A	12 clear tips	8	204	0.35	7.1	1327	0.74595 ± 0.00403	0.09144 ± 0.00044	0.61
	B	20 blocky frags	20	685	0.19	15.8	5138	0.80066 ± 0.00204	0.09457 ± 0.00022	0.92
	C	13 clear tips	10	750	0.33	21.2	2056	0.75231 ± 0.00207	0.09193 ± 0.00021	0.76
	D	2 prismatic needles	1	553	0.09	42.1	102	0.91076 ± 0.02896	0.10237 ± 0.00064	0.21
Øksfjord Gabbro RJR02-3B	A	Numerous clear frags	17	411	0.88	2.7	14553	0.73676 ± 0.00215	0.09069 ± 0.00026	0.91
	B	15 cracked brown frags	14	1421	0.72	2.9	38614	0.74052 ± 0.00210	0.09102 ± 0.00024	0.96
	C	Numerous brown frags	164	399	0.85	16.0	22974	0.72582 ± 0.00246	0.08959 ± 0.00029	0.98
	D	12 clear euhedral	15	416	0.70	23.5	1515	0.72935 ± 0.00314	0.09001 ± 0.00046	0.65
	E	10 blocky clear frags	9	2025	0.72	65.8	1567	0.72659 ± 0.00278	0.08935 ± 0.00030	0.74
Øksfjord Monzonite RJR02-41C	A	Numerous clear euhedral	148	870	1.05	7.0	103588	0.73420 ± 0.00445	0.09030 ± 0.00054	0.99
	B	Numerous clear frags	217	483	0.86	24.6	24441	0.74231 ± 0.00402	0.09132 ± 0.00048	0.99
	C	Numerous clear euhedral	213	974	1.22	43.6	27010	0.73527 ± 0.00291	0.09055 ± 0.00034	0.98
	D	Numerous clear euhedral	264	1390	1.43	34.2	61242	0.73881 ± 0.00234	0.09100 ± 0.00027	0.98
Øksfjord Monzodiorite RJR02-40B	A	Numerous clear euhedral	327	143	0.54	14.6	18171	0.73602 ± 0.00273	0.09064 ± 0.00032	0.97
	B	Numerous brown tips	239	59	0.41	5.8	13713	0.73691 ± 0.00197	0.09070 ± 0.00022	0.93
	C	Numerous brown frags	58	137	0.62	6.3	7192	0.73919 ± 0.00307	0.09110 ± 0.00040	0.79
	D	Numerous blocky frags	35	80	0.43	13.4	1208	0.73346 ± 0.00278	0.09042 ± 0.00025	0.68
	E	Numerous clear euhedral	21	160	0.56	3.5	5462	0.73682 ± 0.00442	0.09077 ± 0.00054	0.91
Øksfjord Norite RJR03-129C	A	3 clear euhedral	19	1032	0.77	7.6	14752	0.74300 ± 0.00260	0.09125 ± 0.00032	0.94
	B	7 clear euhedral	15	1048	0.77	7.3	12162	0.73230 ± 0.00280	0.09011 ± 0.00034	0.94
	C	9 clear euhedral	26	1706	0.76	30.5	8234	0.73300 ± 0.00390	0.09014 ± 0.00047	0.98
	D	10 clear frags	23	1722	0.67	49.4	4159	0.72730 ± 0.00470	0.08960 ± 0.00057	0.97
Øksfjord Pyroxenite RJR03-129D	A	15 brown euhedral	10	418	0.56	14.0	1724	0.74358 ± 0.00282	0.09161 ± 0.00028	0.69
	B	Numerous clear frags	252	428	0.63	29.3	21149	0.74570 ± 0.00261	0.09170 ± 0.00031	0.98
	C	Numerous clear frags	25	595	0.72	11.1	7770	0.74790 ± 0.00228	0.09195 ± 0.00026	0.95
Øksfjord Granite RJR03-129A	A	12 clear euhedral	22	616	7.46	9.7	8337	0.78020 ± 0.00420	0.09462 ± 0.00053	0.91
	B	4 clear euhedral	13	697	0.25	9.7	5310	0.73750 ± 0.00260	0.09061 ± 0.00032	0.90
	C	3 prismatic needles	3	1264	2.11	3.0	7436	0.76430 ± 0.00490	0.09315 ± 0.00058	0.96
	D	11 brown cracked	29	979	0.25	12.7	13058	0.76250 ± 0.00220	0.09291 ± 0.00028	0.87
	Mon1	Monazite	1	2787	23.46	60.1	282	0.73850 ± 0.00990	0.09114 ± 0.00054	0.46
	Mon2	Monazite	1	919	23.13	350.3	33	0.73410 ± 0.14740	0.09072 ± 0.00236	0.11

Footnotes

- a** Model value calculated from ²⁰⁸Pb/²⁰⁶Pb ratio and the age of the sample
b Total common lead, including analytical blank and initial common lead in the sample
c Corrected for spike contribution and fractionation
d Corrected for spike contribution, fractionation, blank and initial common lead (as calculated from Stacey and Kramers, 1975), errors reported at 2σ
e 2σ absolute errors reported in Ma

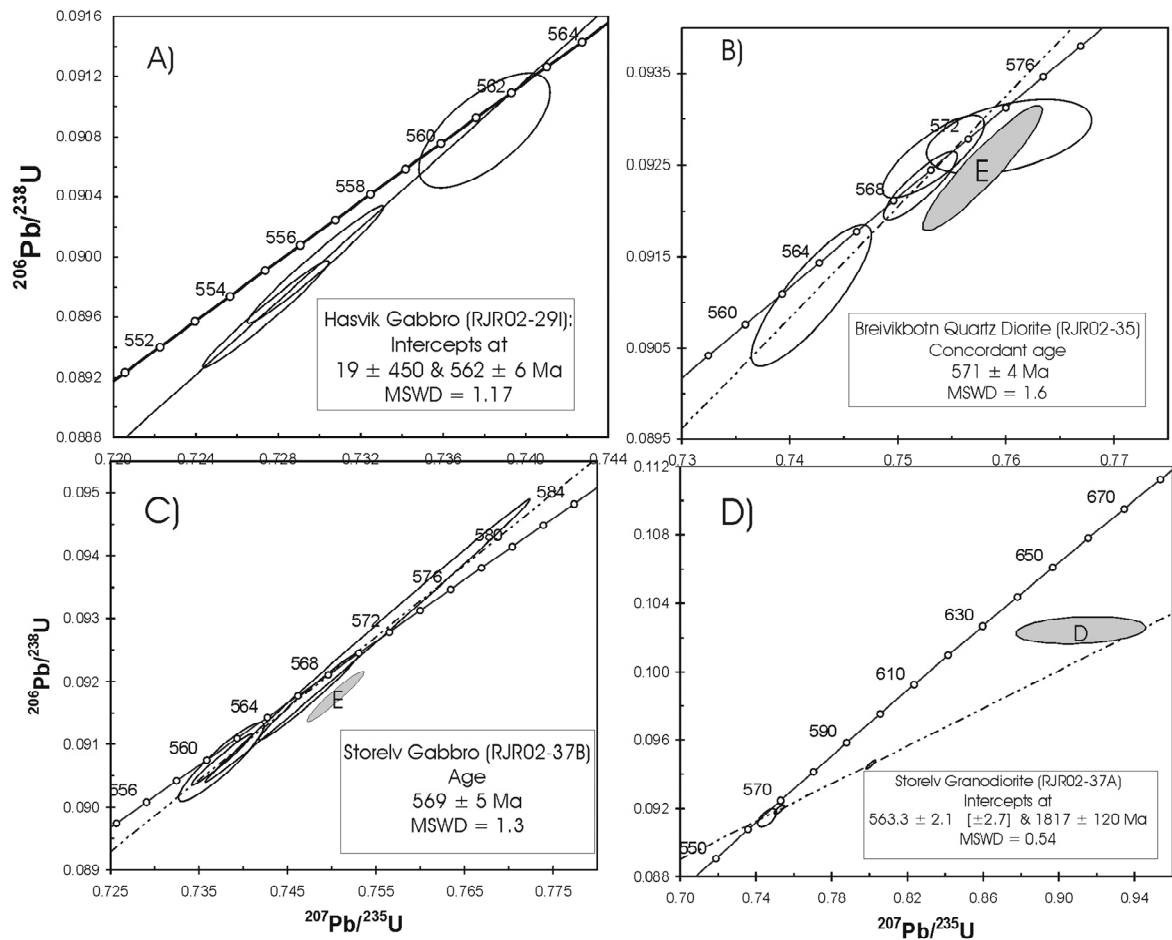


Figure 4.1- Concordia diagrams for TIMS U/Pb analyses from rock samples from the island of Sørøy, plotted from data given in Table 4.1. Errors are 2σ except where noted. a) Hasvik gabbro RJR02-29I - unanchored regression. b) Quartz diorite RJR02-35 from Breivikbotn - analysis E is discarded for analytical reasons, whereas analysis D is not used to derive the concordant age. Unanchored regression. c) Storelv Gabbro RJR02-37B- analysis E is discarded for analytical reasons, and a broad anchoring point of 200 ± 200 Ma is used to calculate the age. d) Storelv Granodiorite RJR02-37A - analysis D discarded for analytical reasons, unanchored regression. Further details on the interpretation of these data may be found in the text.

For RJR02-29I, several grains and some fractions were analysed. However, all of the analyses are reversely discordant, reflecting either undetected analytical problems or open system behaviour with Pb gain or U loss, and causing uncertainty in the interpretation. For this reason the data cannot yield a good age and are not used in Fig. 4.1. However, in light of the age extracted for RJR02-29I, it can be inferred that the reverse discordance was probably due to some U-Pb fractionation process. Three

fractions of large zircon grains (80-150 μg) from RJR02-29I yield concordant to slightly discordant data forming a linear array with an upper intercept age of 562 ± 6 Ma (Figure 4.1). Since the zircons are generally concordant and do not show either significant alteration or a younger component on the Concordia, it is likely that the zircons are magmatic in origin.

4.2.2. The Breivikbotn Gabbro and Diorite, Sørøy

The Breivikbotn Gabbro is a difficult target for U/Pb zircon dating, as it is generally very mafic in composition and thus zircon-poor. There are pegmatitic horizons of various compositions within the gabbro, but distinguishing late-stage igneous pegmatites from the numerous alkaline intrusions or pegmatites developed in small shear zones proved difficult. Instead, a tonalite intruded along the eastern edge of the gabbro was investigated. Although poorly exposed, the unit has been mapped by Sturt & Ramsay (1965), who considered it to be genetically related to the mafic pluton, as both bodies have been deformed and metamorphosed by the same deformation events. However, it is possible that the tonalite is significantly younger than the gabbro, so the tonalite only provides a lower limit on the age of the Breivikbotn Gabbro.

A 1 kg sample of tonalite (RJR02-35; Table 4.1) yielded nearly 20 milligrams of igneous zircon. Most zircons are brown and cracked, but some crystal shape could be discerned. Five fractions were picked for analysis, composed mainly of whole grains or broken tips from elongate grains. Of these, one analysis (RJR02-35 E) was conducted on the secondary electron multiplier at low temperatures, and can be considered a poor analysis for technical reasons and excluded from the age calculations. One of the other analyses (RJR02-35 D) is not concordant, but does plot

on a linear regression line (Fig. 4.1) with the other three analyses at an age of 571 ± 4 Ma (MSWD= 1.6). This is considered the age of formation of the tonalite.

4.2.3. The Storelv Gabbro and associated Granitoid, Sørøy

The Storelv Gabbro is described in Chapter 2. Zircon analyses obtained from samples of Storelv Gabbro (RJR02-37B) and the associated granodiorite (RJR02-37A) are presented in Table 4.1. Twenty grains of zircon were retrieved from the gabbro (RJR02-37B). The zircons were brown, and generally display rounded edges but euhedral crystal shapes. Five single grains were analysed. Of these, one analysis (RJR02-37B E) is considered suspect, as the $^{207}\text{Pb}/^{206}\text{Pb}$ ratio measured on the secondary electron multiplier showed significant variation during the analytical run. The remaining analyses, all clustered near Concordia, show some dispersion in U-Pb, suggestive of mild Pb-loss (Fig. 4.1c). A Discordia line anchored at 200 ± 200 Ma, to cover all possible times of Pb-loss, yields an upper intercept age of 569 ± 5 Ma (MSWD = 1.3). These zircons show no signs of an older isotopic component, and are most likely igneous in origin.

The granodiorite provided a large amount of zircon (>1 g). However, granitic melts commonly contain inherited zircons, and this granodiorite is no exception. As inherited cores could not be discerned under the microscope, several small fractions were analysed in order to check for inheritance. In Figure 4.1d it can clearly be seen that most of the analyses are discordant, but that they do form a linear array. All four fractions yield a regression line with intercepts at 559 ± 18 Ma and 1452 ± 310 Ma (MSWD = 2.3). The three lowermost points alone (omitting point D, which contains much common Pb and whose cores may have a separate source), have a better fit (MSWD = 0.5), and define a more precise lower intercept age of 563 ± 2 Ma,

representing the age of magmatic crystallisation of the rock, whereas the upper intercept age of 1817 ± 130 Ma points to a Palaeoproterozoic source for the xenocrystic component. The 563 ± 2 Ma age of the granodiorite is within error of the age of its host gabbro (569 ± 5 Ma), but field observations suggest it is somewhat younger.

4.2.4. Mafic rocks from the Øksfjord Peninsula

RJR-02-3, as described in Chapter 3, is an example of the gabbroic host for the numerous monzonitic and syenitic sheets along the east coast of Øksfjord. The zircons retrieved from this gabbro are generally fractured and brown, although there is a population of clear zircons as well. There was some indication of resorption around the edges of some crystals, but some crystal shape could be discerned in most crystals. Five analyses are presented in Table 4.1. These data do not overlap each other on the Concordia. It is likely that this is partly due to instrumental error, but the generally high U content of the samples also indicates that Pb-loss is likely to have affected some of the crystals. In order to utilise the analyses in calculating an age for the rock, the analytical errors were doubled and a broad anchoring point (200 ± 200 Ma) used to accommodate the possible Pb loss. The resulting regression (Fig. 4.2a) yields an age of 565 ± 9 Ma (MSWD = 2).

Figures 4.2b and 4.2c show the Concordia diagrams for the samples RJR-02-40B and RJR-02-41C. In both cases, a broad anchoring point at 200 ± 200 Ma was used to accommodate any potential Pb-loss. As was the case for the gabbro, the data for RJR02-41C are clustered close to the Concordia but there is some excess scatter, which probably has to be attributed to an underestimated analytical uncertainty.

Calculation using double errors yields an age of 566 ± 4 Ma (MSWD = 1.6). A regression through the data for RJR02-40B has a good fit and yields an age of 565 ± 5 Ma (MSWD = 0.54). All three ages for these rocks overlap within error, and the ages are all considered to be primary (see below).

Of the rocks taken from the west coast of Øksfjord, both RJR03-129C and D provided significantly more zircon than is normally expected from mafic rocks, and thin section examination shows that both also contain apatite as an accessory phase. It is likely that both crystallised from crustally-contaminated magmas, similar to the situation with the Hasvik Gabbro (Tegner et al., 1999). The zircon analyses are again presented in Table 4.1.

Four zircon fractions were analysed from RJR03-129C (Fig. 4.2d). These fractions form a linear array, with a very tight spread. Unfortunately, none of the zircons is concordant. In this case, the choice of anchoring point influences the resulting age regression considerably. With no anchoring point, the regression line has intercepts at 580 ± 24 Ma and 366 ± 220 Ma (MSWD = 0.3), possibly indicating early Pb-loss. A more precise age of 569 ± 9 Ma is obtained by using an anchoring point of 200 ± 200 Ma (MSWD = 0.94), and this is the preferred age. Three fractions were analysed from RJR03-129D (Fig. 4.2e). One analysis plots above Concordia, and is excluded. An unanchored age regression yields an age of 566 ± 1 Ma (MSWD = 0.11).

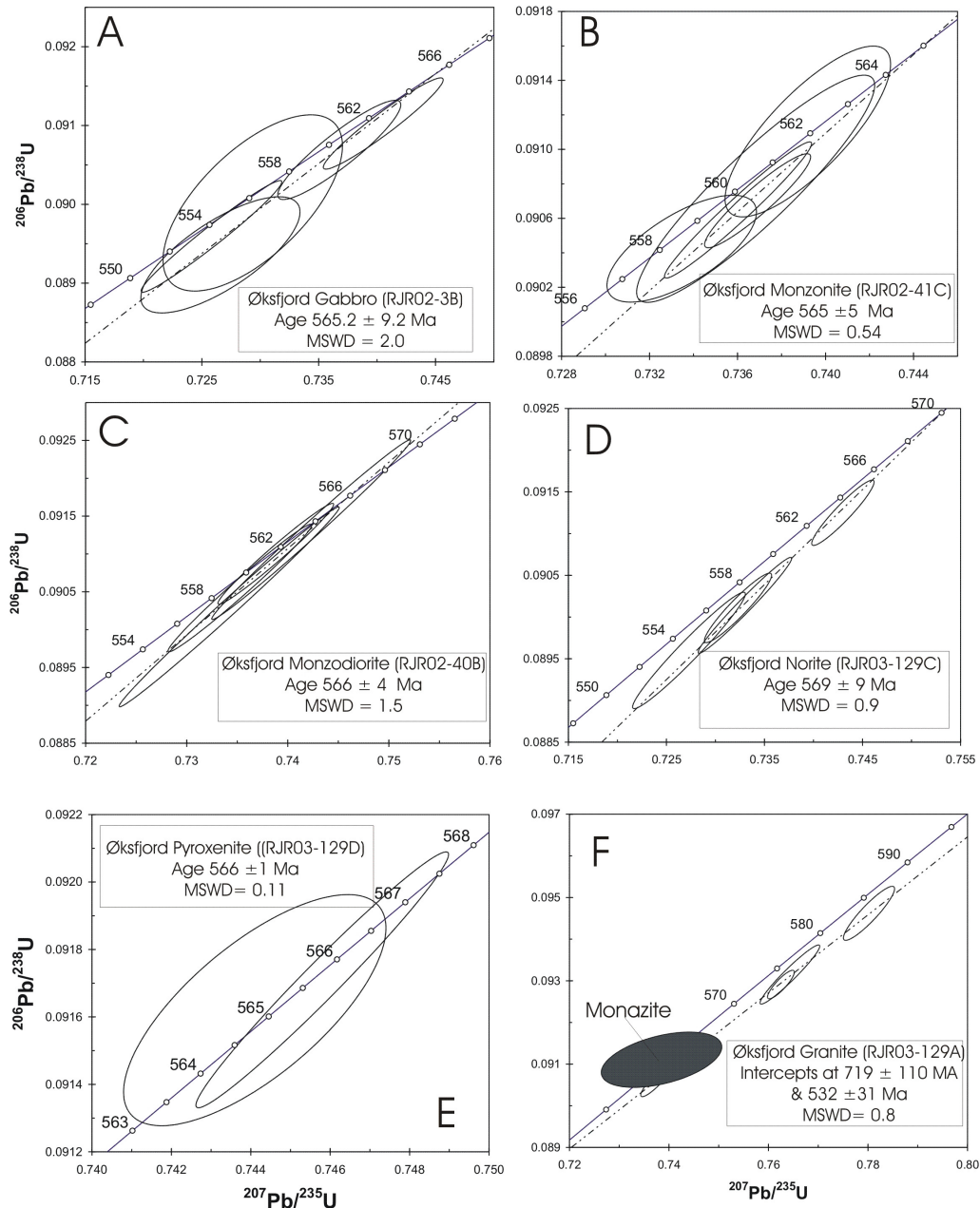


Figure 4.2- Concordia diagrams for TIMS U/Pb analyses from rocks from the Øksfjord peninsula, plotted from data given in Table 4.1. Errors are 2σ except where noted a) Gabbro RJR02-3B - an age was obtained by doubling the analytical uncertainty on each analysis to correct for poor instrumentation analysis, and using an anchoring point of 200 ± 200 Ma. b) Monzonite RJR02-40B - an age was obtained using a broad anchoring point of 200 ± 200 Ma. c) Monzodiorite RJR02-41C - an age was obtained using a broad anchoring point of 200 ± 200 Ma. d) Norite RJR03-129C - an age was obtained using a broad anchoring point of 200 ± 200 Ma. e) Orthopyroxenite RJR03-129D - no anchoring point was required to calculate this age. f) Granite RJR03-129A - displays both monazite and zircon analyses. Further details on the interpretation of these results may be found in the text.

The granite body (RJR03-129A) that intrudes the mafic rocks contains zircons that appear to have been inherited from another source (Fig. 4.2f). The four zircon analyses define a Discordia line between 719 ± 110 Ma, and 532 ± 31 Ma (MSWD = 0.8) whereas one analysis of monazite has a $^{207}\text{Pb}/^{235}\text{U}$ age of 561 ± 4 Ma (in this case virtually identical to the Concordia age of 562 ± 3 Ma). Although monazite can occur as an inherited phase in granitic rock, the age determined from this sample overlaps with other ages in the area and is not considered coincidental. The upper intercept age of this line (719 ± 110 Ma) points to a provenance of the melt from Neoproterozoic crust, broadly consistent with more recent, unpublished results collected during this project.

4.2.5. The Breivikbotn Carbonatite, Sørøy

Two rock samples from the Breivikbotn Carbonatite supplied minerals for U-Pb TIMS analysis. These rocks (samples RJR02-34D and E; Table 4.2) are coarse-grained carbonatitic rocks. Zircon occurs as crystals of up to mm in size in sample 34D and could easily be extracted from the rock. Sample 34E is rich in carbonate and was partially dissolved in HCl, liberating the silicates, including titanite and some zircon. One large zircon fragment from sample E was abraded vigorously, breaking it in pieces that were used for the three analyses given in Table 4.1. Conversely, several fragments from sample D were abraded and analysed in two separate fractions. The combined data (Figure 4.4) reveal relatively high U contents, variable Th/U and are concordant to slightly discordant yielding a discordia line with intercept ages of 574 ± 5 and 372 ± 98 Ma (MSWD = 1.7). These ages are interpreted as indicating the time of magmatic crystallization and (within error) the time of Scandian metamorphism, respectively. Titanite was analyzed in sample 34E. It has low U contents (10-20 ppm),

relatively high Th/U (5-6) and a relatively low initial Pb content (0.4-0.7). The three titanite analyses have poor precision, but they yield $^{206}\text{Pb}/^{238}\text{U}$ ages of 568-558 Ma, which are consistent with the magmatic crystallisation and also indicate that they were only mildly reset during the Scandian event.

Table 4.2: U/Pb analyses of rocks from the Breivikbotn Alkaline Complex

Fraction Analysed	Weight [ug]	U [ppm]	Th/U ^a	Pb com ^b [pg]	$^{206}\text{Pb}/^{206}\text{Pb}^c$	$^{207}\text{Pb}/^{235}\text{U}^d$	2 sigma [abs]	$^{206}\text{Pb}/^{238}\text{U}^e$	2 sigma [abs]	rho	$^{207}\text{Pb}/^{206}\text{Pb}^d$	2 sigma [abs]	$^{207}\text{Pb}/^{206}\text{Pb}$ Age (Ma) ^e	2 sigma [abs]
RJR-02-34D shonkinite														
81/15 z fr cl 40gr	67	246	2.2	4.0	24014	0.7517	0.0024	0.09222	0.00028	0.95	0.05912	0.00006	571.4	2.2
81/16 z cl (-lu) cr	100	366	2.4	5.5	38204	0.7510	0.0024	0.09213	0.00028	0.97	0.05912	0.00005	571.6	1.8
RJR-02-34E carbonate-rich shonkinite														
81/21 z fr cl 1gr	38	388	1.0	3.4	25479	0.7556	0.0020	0.09271	0.00023	0.95	0.05911	0.00005	571.0	1.8
84/52 z fr cl 1gr	67	273	0.3	1.9	53813	0.7405	0.0017	0.09105	0.00019	0.95	0.05899	0.00004	566.7	1.5
81/20 z fr cl 18gr	34	307	0.6	5.2	11398	0.7363	0.0020	0.09065	0.00023	0.94	0.05891	0.00005	563.9	2.0
84/S.50 tit br-y 13gr	92	12	6.2	58.9	126	0.7291	0.0205	0.09218	0.00048	0.17	0.0574	0.0016	506	60
80/S.52 tit fr (>50)	75	22	5.6	57.7	180	0.7299	0.0140	0.09041	0.00040	0.25	0.0586	0.0011	551	40
84/S.51 tit br-y (in) 11gr	96	20	5.8	41.7	278	0.7262	0.0092	0.09039	0.00033	0.35	0.05826	0.00069	540	26
RJR-03-116 syenite gneiss														
95/26 z cl fr 10gr	18	87	20.0	1.4	6420	0.7541	0.0031	0.09262	0.00035	0.86	0.05905	0.00012	568.8	4.5
95/25 z cl fr 40gr	1	151	4.6	0.4	2506	0.7505	0.0039	0.09231	0.00040	0.79	0.05897	0.00019	565.8	7.0
66/62 z fr	50	114	1.7	5.7	5788	0.7504	0.0021	0.09220	0.00026	0.82	0.05903	0.00010	568.2	3.6
75/60 z fr	367	379	5.0	137.9	5750	0.7367	0.0020	0.09079	0.00022	0.96	0.05885	0.00006	561.7	1.8
75/58 z fr	349	11	61.2	42.8	529	0.7296	0.0046	0.09030	0.00028	0.46	0.05860	0.00034	552	12
75/61 z fr	388	16	28.3	333.1	121	0.7327	0.0194	0.09043	0.00037	0.09	0.05878	0.00156	559	58
66/60 z fr	247	246	6.6	99.4	3331	0.6958	0.0018	0.08648	0.00018	0.90	0.05835	0.00007	543.1	2.5
95/S.18 X br 1gr	6	111722	0.1	848.2	3323	0.5084	0.0022	0.06677	0.00027	0.96	0.05522	0.00006	421.2	2.5
72/61 z tips	167	19	35.7	136.9	116	0.5059	0.0149	0.06643	0.00035	0.10	0.0552	0.0016	422	64
72/57 z fr	195	6	265.5	87.8	83.9	0.6467	0.0327	0.08465	0.00064	0.19	0.0554	0.0028	429	----
66/52 z fr	202	9	105.7	28.7	390	0.7113	0.0155	0.09175	0.00086	0.39	0.0562	0.0011	461	44
75/62 z fr	475	12	49.5	137.1	243	0.7100	0.0100	0.08775	0.00023	0.24	0.05868	0.00080	555	30
75/59 z fr	283	102	6.6	146.5	1073	0.6882	0.0025	0.08531	0.00018	0.64	0.05851	0.00016	548.9	6.1
66/57 z fr	632	24	38.7	45.0	1808	0.6773	0.0034	0.08474	0.00036	0.84	0.05797	0.00016	528.7	6.0
72/62 z tips	72	94	36.4	45.2	368	0.2608	0.0061	0.03740	0.00031	0.45	0.05058	0.00106	222	48
RJR-04-236 nepheline syenite dyke														
93/12 z fr cl 1gr	1	117	1.5	3.0	241	0.740	0.015	0.09119	0.00043	0.42	0.05885	0.00110	561	40
93/11 z fr cl 1gr	1	99	4.3	1.5	392	0.731	0.013	0.08998	0.00047	0.43	0.05894	0.00095	565	35
RJR-04-245 alkaline gneiss														
98/3 z cl fr 18gr	3	480	2.1	2.2	3831	0.7502	0.0028	0.09208	0.00026	0.85	0.05909	0.00012	570.4	4.3
108/59 z cl fr 1gr	16	11	42.6	1.8	587	0.7500	0.0088	0.09187	0.00038	0.49	0.05921	0.00062	575	22
108/57 z cl fr 30gr	1	163	4.0	1.5	632	0.7165	0.0081	0.08808	0.00041	0.53	0.05900	0.00057	567	21
98/6 z eu-sb sp 7gr	7	100	1.1	7.6	450	0.5883	0.0047	0.07514	0.00022	0.46	0.05678	0.00041	483	16
108/58 z an eq-el 11gr	6	344	0.7	3.5	2653	0.5562	0.0017	0.07187	0.00016	0.74	0.05613	0.00011	457.4	4.5
RJR-04-246 syenite dyke														
98/8 z fr plbr 1gr	17	37	6.2	2.1	1709	0.7418	0.0075	0.09125	0.00085	0.93	0.05896	0.00021	565.6	7.8
98/9 z an(-sb) eq 30gr	11	113	2.8	2.4	2979	0.7367	0.0022	0.09054	0.00021	0.74	0.05901	0.00012	567.5	4.4
98/7 z fr cl 7gr	35	107	1.5	5.8	3606	0.7160	0.0018	0.08868	0.00020	0.84	0.05855	0.00008	550.6	3.0
98/S.2 m an y tr NA 1gr	2	11	954.5	3.3	46.9	0.46	0.10	0.0691	0.0011	0.66	0.048	0.010	106.5	---

Key for fraction analysed

= zircon; X = unidentified brown mineral (xenotime?); m = monazite, fr = fragment, eq = equant, an = anhedral, eu = euhedral, sb = subhedral, tr = transparent, cl = clear, colourless; y = yellow; br = brown; plbr = pale-brown, incl = inclusions, NA = non-abraded

Footnotes

- Model value calculated from $^{208}\text{Pb}/^{206}\text{Pb}$ ratio and the age of the sample
- Total common lead, including analytical blank and initial common lead in the sample
- Corrected for spike contribution and fractionation
- Corrected for spike contribution, fractionation, blank and initial common lead (as calculated from Stacey and Kramers, 1975), errors reported at 2 σ
- 2 σ absolute errors reported in Ma

4.2.6. The Breivikbotn Syenitic Complex, Sørøy

Several samples were taken from the Breivikbotn Syenitic Complex, as detailed in Chapter 2. Whereas RJR03-115 proved barren of zircons, RJR03-116 contains much zircon, up to 1-2 mm in size and visible in hand specimen as brown pyramidal crystals surrounded by a brown halo of radiation damage. After crushing, the population

comprised small, clear zircons with euhedral shape but generally rounded edges, and large (100-1000 μm) brown fragments. In general, most grains are somewhat turbid and SEM analyses of some of the large zircon grains found them to be riddled with radioactive minerals, including xenotime and a thorium-bearing silicate (Figure 4.3). Fractions of the smaller zircons and single grains from the larger zircons were analysed (Table 4.1). Initial results revealed a wide spread of ages, some of the analyses plotting concordantly at 570, 420 and 230 Ma and several reversely discordant (Figure 4.4). Subsequent analyses, focused on smaller crystals and fragments with no, or a minimum of inclusions and alteration, reduced the amount of discordance and the scatter. Overall the analyses also show a considerable variation in the content of U (6-379 ppm), Th/U (1.7 – 265 ppm) and common Pb (0 – 0.85 ppm). In general high values of Th/U and common Pb are a characteristic of the fractions showing the most irregular isotopic behaviour. These indications combined with the SEM study suggest that the discordance of the zircon must have been controlled to a large extent by the behaviour of the inclusions and by alteration of metamict domains.

As a test one small unidentified brown mineral, similar to some of those visible inside zircon, was analyzed. The result shows a high U content of over 10%, high common Pb of 140 ppm, but low Th/U. It could possibly represent xenotime, such as those observed by SEM within zircon. The analysis is slightly discordant, plotting close to 420 Ma, the age of the Scandian metamorphic event, supporting a hypothesis in which resetting of the zircon occurs mainly by Pb-loss from U and/or Th-rich inclusions and altered domains. It is not evident what the basic concentration of U and Th/U is in the zircon itself. Even the grains which contain no inclusions visible under a high-magnification binocular microscope and those zircons with the uppermost concordant

analyses show strong variations in these elements, including U contents that are considerably higher than those in more discordant analyses. It appears thus likely that the variations in Th and U are at least in part a characteristic of the zircons themselves. High Th/U ratios with locally extremely low U contents are typical features of zircon in alkalic rocks, especially in carbonatites (e.g., Amelin and Zaitsev 2002; Dahlgren, unpublished data).

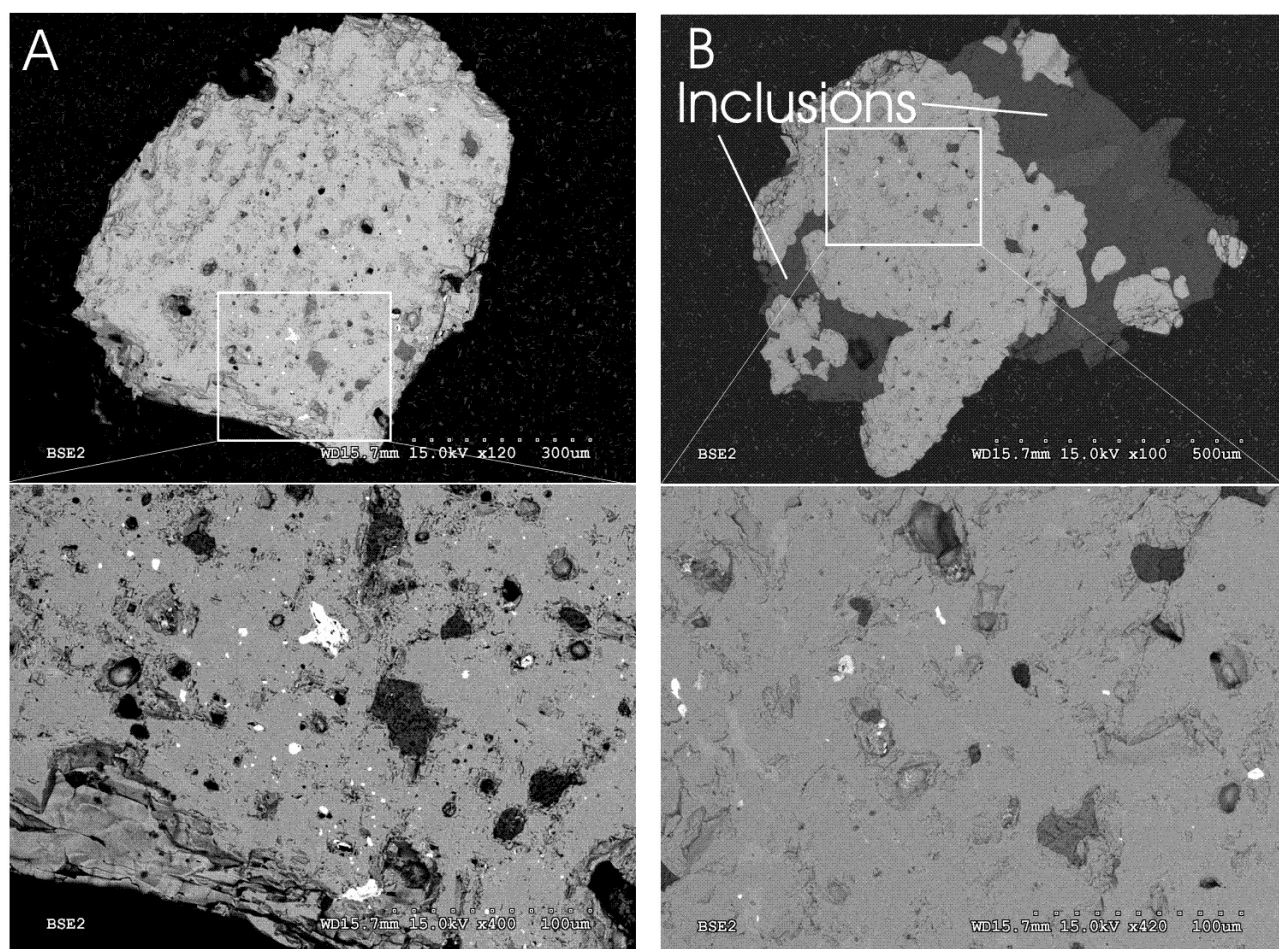


Figure 4.3- Back-scattered electron images of zircons from alkaline gneiss RJR03-116. A) Zircon showing numerous small inclusions. Bright spots are generally thorite, darker inclusions may be xenotime or apatite. B) Zircon intergrown with xenotime. Magnified area shows numerous small inclusions.

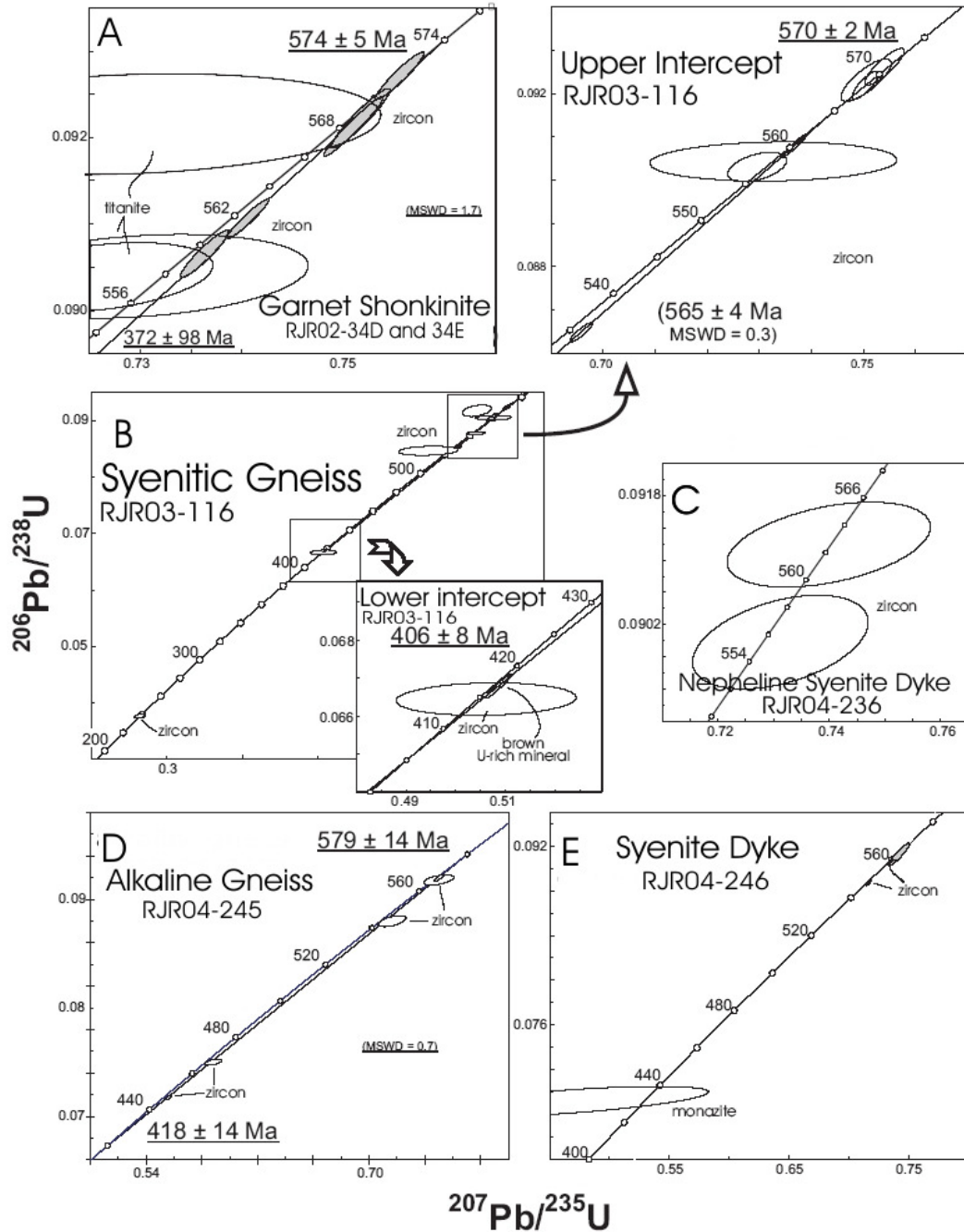


Figure 4.4 - Concordia diagrams for rocks analysed in the Breivikbotn area of Sørøy. All errors are 2σ , and the results are further discussed in the text. A) Garnet shonkinite from the Breivikbotn Carbonatite, using analyses from samples RJR02-34D and 34E. B) Nepheline syenite RJR03-116. Owing to the complexities in the analyses of the zircons from this rock, the diagram is further broken down to show the upper and lower intercepts of the Concordia plot. C) Syenitic dyke RJR04-236. No determination of absolute age was made from these concordant zircons, owing to the spread between the analyses, although a relative age can be estimated. D) Alkaline gneiss RJR04-245, host to syenite dyke RJR04-245. E) Syenite dyke RJR04-246, hosted by alkaline gneiss RJR04-245. No determination of absolute age was made from these concordant zircons, owing to the spread between the analyses, although a relative age can be estimated.

Several of the zircon analyses in RJR-03-116 (Table 4.1, Fig. 4.3) show a highly disturbed data pattern, for example the analysis plotting at 237 Ma, and those that are reversely discordant. Their behaviour is probably due to a number of effects, including partial metamictization and especially loss or gain of Pb and U from the various inclusions. The least discordant analyses show instead a more coherent behaviour. The six uppermost analyses are co-linear (MSWD = 0.4) defining an upper intercept age of 564 ± 5 Ma. Inclusion in the regression of the analysis for the brown high-U mineral and the overlapping zircon data point refines the parameters (MSWD = 0.3 and 565 ± 4 Ma) improving the precision of the lower intercept age to 406 ± 8 Ma. In spite of the good colinearity, it is evident that the upper intercept age is somewhat younger than the age of the three uppermost data points alone, possibly reflecting a faint shift due to the complex isotopic behaviour of the population, perhaps with an added ^{230}Th disequilibrium effect. The three uppermost data points span an age range from 568 to 571 Ma and are considered to yield the most reliable indication for the age of the rock, which on this basis is estimated at 570 ± 2 Ma. The age overlaps those for plutonic rocks in the area, including the Breivikbotn Carbonatite. The lower intercept age of 406 ± 8 Ma overlaps with the lower intercept of the Breivikbotn Carbonatite, and is a metamorphic age reflecting the Scandian event at 420 Ma (Roberts, 2003).

Close to this location, a nepheline syenite dyke, crosscutting both alkaline gneiss similar to RJR03-116 and the host gabbro, was sampled for dating (RJR-04-236). This dyke is 40 cm wide and shows an alignment of biotite crystals parallel to the walls of the dyke. The dyke coarsens towards the middle, with biotite being concentrated along the sides. Zircon is scarce in this sample and consists mainly of turbid grains

similar to those in RJR-03-116. There is also abundant monazite, but an attempt to analyze one grain showed a very low U content and almost no radiogenic Pb. A selection of zircon fragments with a minimum of inclusions was abraded strongly and the two best, inclusion-free grains produced by this process were analyzed. They have moderate U contents (ca. 100 ppm), relatively high Th/U (1.5-4.3) and define concordant but not overlapping analyses suggesting some loss of Pb (Table 4.2, Fig. 4.3). By themselves the data do not define a precise age, but provide a minimum age of 563 ± 3 Ma, which is consistent with the age of the other alkalalic, gabbroic and granitic rocks in the area.

At Hvitnes, numerous alkaline dykes intrude the Klubben Psammite. Many pictures of this locality can be found in Sturt & Ramsay (1965). At one location, deformed alkaline gneiss (RJR-04-245) and country rock are crosscut by a thin (0.4 m), apparently undeformed dyke of biotite-free syenite (RJR-04-246). Both samples contain comparable zircon populations, also very similar to those in the alkalalic gneiss and dyke discussed above. The zircons occur mainly as anhedral fragments and as small euhedral to subhedral crystals. Inclusions are ubiquitous and most grains tend to be turbid. Selections of small crystals or of clear fragments free of inclusions were strongly abraded and some of the resulting clean grains used for analysis. The obtained five zircon data points are spread out along a discordia line with intercepts at 579 ± 14 Ma and 418 ± 14 Ma (MSWD = 0.7; Figure 4.4). Data for the fragments plot near the top of the line whereas the small subhedral crystals define the lower end of the line. The distinction, in terms of discordance, between the two types of zircon suggests that the small grains are metamorphic newly-grown or recrystallized zircon. They show about the same, moderate but quite variable U-contents as the large,

possibly primary, fragments and lower Th/U ratios, although the Th/U variation within the fragments is enormous. Three zircon fractions were analysed for sample RJR-04-246. They plot in the 550-570 Ma segment of the Concordia diagram and in this case there is no distinction between fragments and small anhedral zircons. As for RJR-03-236 the three analyses do not yield a precise age but their position and characteristics leave no doubts that they are coeval and probably cogenetic products of the same alkaline magmatic event. Sample RJR-04-246 also contains monazite. One crystal was analysed showing a U content of only 11 ppm in contrast to about 1% Th, and although the precision of the analysis is very low, it indicates a probable formation during the Scandian metamorphism.

4.3. Discussion

4.3.1. Comparisons with previous isotopic work

The ages in this study differ significantly from some of the other published ages in the area, and this needs to be addressed. Figure 4.5 compares the spread of ages from previous work with the ages obtained in this study. In general, most previous age determinations overlap within error of the U/Pb results reported in this study. The better accuracy and precision of the U/Pb analyses supersedes these earlier results, and the age spread of the U/Pb results is taken to represent the age of intrusion of the SIP gabbros. However, several analyses, using both Rb-Sr and Sm-Nd techniques, fall well outside the range of ages delimited during this study.

The Sm-Nd and Rb-Sr isotopic dating systems can both yield spurious results, for various reasons. The Rb-Sr isotopic system may remain open, especially in rocks undergoing deformation. Thus, the Rb-Sr system is often reset, producing ages that

are younger than the actual formation ages of the rocks. The Sm-Nd system is much more resistant to resetting but can be affected by isotopic disequilibrium. If suites of coexisting minerals or whole rocks were not in isotopic equilibrium with each other, such as occurs occasionally in a layered intrusion, it is possible to obtain ages that are older or younger than the true age of the rock. This would require fractionation and variable assimilation operating simultaneously in the melt to create disequilibrium, and a subsequent mechanism to prevent diffusion and re-equilibration as the melt cools and solidifies. This effect has been reported from many cases, for example

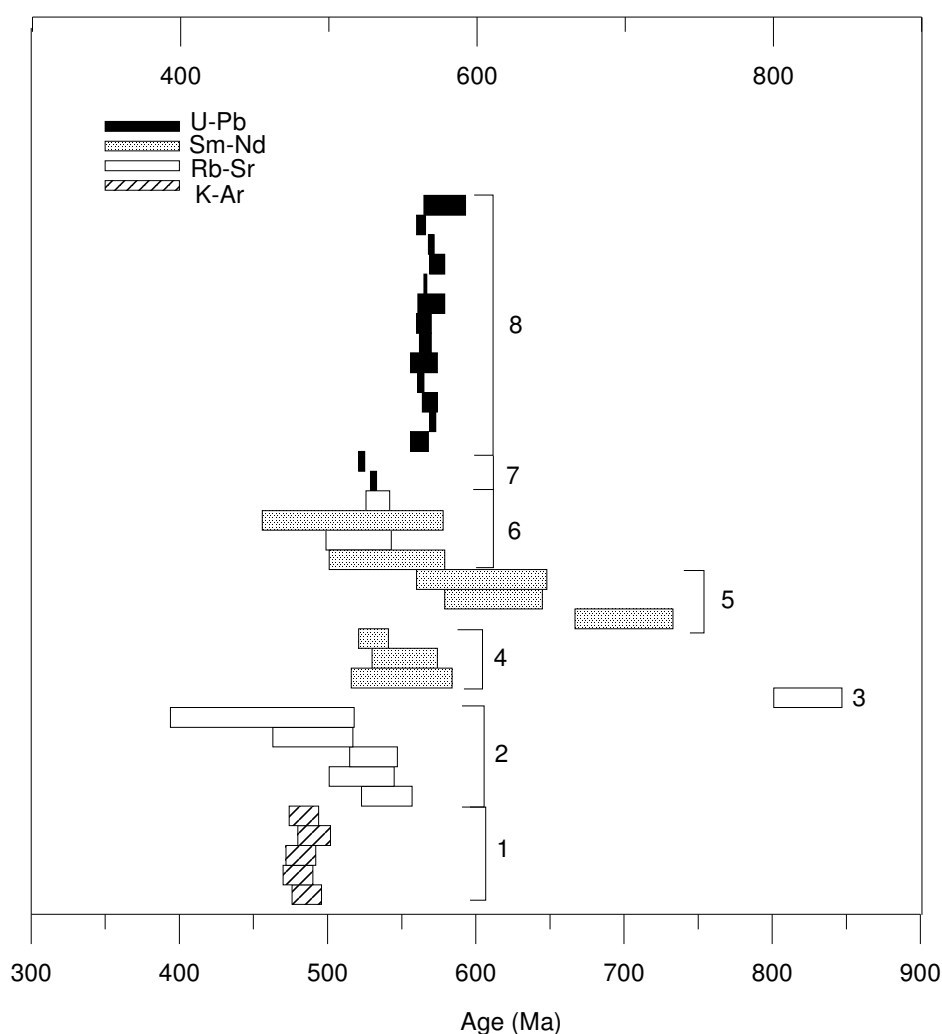


Figure 4.5: Compilation of published ages and dating methods from the Seiland Igneous Province. References: 1= Sturt et al. (1967), 2= Sturt et al. (1978), 3= Krogh & Elvevold (1991), 4= Mørk & Stabel (1990), 5= Daly et al. (1991), 6= Cadow (1993), 7= Pedersen et al. (1989), 8= this study.

Proterozoic anorthosites (Ashwal & Wiebe, 1989) and the Bushveld Complex (Kruger, 1994; Prevec et al., 2005). If a whole-rock isochron is used, then great care must be taken that all rocks utilised for the isochron are coeval and do not represent different degrees of crustal contamination, or else the isochron will yield a meaningless age.

Because several radiometric dates within the SIP vary greatly from one isotopic system to another, it is likely that at least one of these effects has biased previous age investigations of the mafic rocks of the SIP. Mixing of unrelated rocks may explain some of the biased data obtained in at least two of the previous studies. The Rb-Sr studies by Brueckner (1973) and Krogh and Elvevold (1991) on intrusions from the Øksfjord peninsula produced whole rock ages of 612 ± 33 Ma and 829 ± 18 Ma respectively, utilising data from distinct rock types ranging from monzonite to peridotite, which may not necessarily have been comagmatic. The ages older than that obtained by U/Pb analysis suggest that the isochrons are rotated away from the true age, probably due to crustal contamination of the more potassic and Rb-rich rocks.

The Sm-Nd age of 700 ± 33 Ma (vs. 562 ± 6 Ma for zircon) reported for the Hasvik Gabbro (Daly et al., 1991) is considered to be invalid for a number of reasons. First of all, zircons recovered from the Hasvik gabbro are generally euhedral, elongate crystals with no obvious cores. This morphology is common in magmatic zircon in mafic rocks, and very unlike that of typical metamorphic zircon. If, in spite of that, the zircon age of the Hasvik gabbro were to record a metamorphic event, one has to wonder about the nature of an event that could totally reset existing zircon, or make new zircon, while leaving the Sm/Nd mineral systems and the main minerals

untouched. Logic dictates that if the Sm/Nd age of 700 Ma was correct, then significant metamorphism must have occurred at 560-570 Ma to grow metamorphic zircon and reset the original igneous age in not only the Hasvik Gabbro, but all other rocks in this study. Considering that zircons obtained from the granitoids at Storelv and Øksfjord still retain much older cores that have not been completely reset, and that none of the other plutons in this study show any indications of an igneous provenance older than 570 Ma, the idea of an older age for the Hasvik Gabbro must be dismissed. Similar arguments can be made for an igneous origin of the other mafic rocks.

One possible reason for the unreasonably old Sm-Nd date is isotopic disequilibrium. It has been suggested that up to 21% of the parental melt of the Hasvik Gabbro was composed of assimilated crustal material (Tegner et al., 1999). It is thus probable that the isochrons combine data for crystals formed early in a less contaminated melt, with crystals formed later from the same but more contaminated melt, yielding a flawed age determination.

4.3.2. The occurrence and composition of zircon in the alkalic rocks from the SIP

In general, dating of the alkalic rocks in this study proved to be rather difficult, even though the zircon abundance in some samples is quite high. The main difficulties are related to (i) the common inclusion of U or Th rich minerals, and (ii) to the localised resetting and growth of new zircon during Scandian metamorphism. The zircons tend to have large variations in U content and Th/U, the latter occasionally reaching values as high as 960 (compared to an average less than 10 in most studies of zircon). Initial common Pb is also very variable, although this value may be controlled by micro-

inclusions. Although it is not always possible to distinguish between Pb-loss and the growth of new metamorphic zircon as an explanation for discordance, at least one of the samples (RJR-04-245) appears from the Concordia plot to have grown new zircon during the metamorphic event. Nepheline syenite RJR-03-116 also displays a bimodal size distribution indicative of new zircon growth, with smaller zircons yielding 420 Ma ages, as opposed to the variation shown by the larger grains and fragments. Sample RJR-04-246 also contains metamorphic monazite with extremely low contents of U, but rich in Th.

The chemical, textural and isotopic complexity of zircon is a testament to its high degree of stability during many geological processes, but it is not immune to alteration. Textural evidence such as blurred primary zoning, convoluted zoning or transgressive recrystallisation has been presented as evidence for the solid state alteration of zircon (Hoskin and Schaltegger, 2003). Dissolution-reprecipitation processes, for which the presence of a fluid is a pre-condition, may also affect zircon, resulting in volume changes, the development of porosity, and the growth of mineral inclusions, or the replacement of the primary mineral by a new phase (Putnis, 2002). Recrystallisation of zircon in a dry environment with no fluid phase has been observed, but is only favoured at ultrahigh temperatures (Möller et al., 2002).

Zircons from the alkaline rocks of the SIP display porosity, irregular and pitted crystal faces, and the presence of xenotime and Th-silicate inclusions, as described above. These features indicate that dissolution-reprecipitation was an important factor influencing zircon stability and growth during metamorphism. Dissolution-reprecipitation processes in zircon and their influence on isotope systematics and

chronology have been studied at a number of localities (Rayner et al., 2005; Tomaschek et al., 2003). Both these studies, and others that describe secondary textures in zircon (i.e. Rubatto, 2002; Schaltegger et al., 1999; Vavra et al., 1996; Vavra et al., 1999), emphasise the high-grade metamorphic conditions prevalent during metamorphism (typically eclogite, granulite and migmatite facies).

These conditions are in contrast to those prevalent in the SIP during Scandian metamorphism, where peak P-T conditions of upper-amphibolite facies (<700°C and 8-10 kbar) have been calculated (Elvevold et al. 1994). Whereas it is not uncommon for amphibolite facies metamorphism to promote zircon growth (Parrish and Noble, 2003), recrystallisation or dissolution-reprecipitation at such conditions is unlikely unless the zircon is metamict (Dempster et al., 2004; Geisler et al., 2003; Rizvanova et al., 2000).

Textural evidence for dissolution-reprecipitation and growth of metamorphic zircon in nepheline syenite metamorphosed at amphibolite facies (600-650°C, 5.5-8.5 kbar) conditions has been reported in the Monashee Complex in southeastern British Columbia (Parrish, 1995; Parrish and Scammell, 1988). The zircons from Monashee have cloudy, well-defined, inclusion-rich, sponge-textured older cores (Crowley and Parrish, 1999), with clear metamorphic rims that grew some 690 Ma later. The growth of the new rims was coincidental with the release of hot fluid from leucogranitic and pegmatite melts crystallising in the vicinity.

Thus, alteration textures in zircon from metamorphosed syenitic rocks in both the Monashee Complex and the SIP indicate dissolution-reprecipitation occurring at

upper amphibolite facies conditions, rather than the higher P-T conditions cited in most cases of zircon alteration. This could point towards the presence of a fluid with properties capable of initiating dissolution-reprecipitation of zircon. Pan and Fleet (1996) showed that rare earth elements and selected high field strength elements, including Zr, are mobile in fluorine-dominated fluids at granulite conditions. However, in the SIP and the Monashee Complex there is little evidence for high F-activity fluids associated with the metamorphism, as neither fluorapatite or fluorite are reported as accessory phases at either locality. It has been shown experimentally that zircon is also susceptible to alteration in a carbonate-rich fluid (Rizvanova et al., 2000), and while CO₂-rich fluids have been reported in the SIP (Elvevold and Andersen, 1993), they were documented in country host-rocks and were related to contact metamorphism during emplacement of the SIP. The reactivation of such fluids during later Scandian metamorphism is not impossible, but the amphibolitisation of the SIP gabbros and the seritisation of the feldspars in the alkaline intrusions strongly indicates a water-dominated fluid during metamorphism of the SIP.

Fluids with high H₂O activity, in the absence of a high F and Cl content and in equilibrium with alkali-syenite rocks, will have a pH > 7, and will have a dissolved ionic component, including K⁺ (Huang and Kiang, 1972). Such fluids could have existed during metamorphism of both the SIP and the Monashee Complex, and Zr mobility has been reported in hydrothermal fluids rich in potassium (Gieré 1990). Such a fluid could thus be the dominant factor in the metamorphism of the SIP zircons. However, these fluids have not affected all minerals in the SIP nepheline syenites, as primary titanite in sample RJR-02 43E is only slightly disturbed by

metamorphism. This highlights the complex nature of the metamorphism associated with the SIP.

4.3.3. The origin of the SIP

The new data reported in this chapter suggest the SIP represents the remains of a short-lived, late Precambrian, igneous intrusive event with a current surface exposure of 5400 km². It is also clear that alkaline magma was being emplaced from early in the magmatic history of the SIP at 570 Ma, and was still being emplaced 50 m.y. later (Pedersen et al., 1989). Considering that only the roots of the igneous plutons have been preserved, it is likely that the original magmatism was considerably more voluminous, and may have included extrusive magmatism, all now long eroded. Rough estimation from the surface outcrop of the different rock types shows the composition of the SIP to be: 50% gabbro, 35% ultramafic, 10% felsic (monzonite, diorite and granitoid), and 5% alkaline (nepheline syenite, syenite, alkaline gneiss, and minor amounts of carbonatite).

The SIP presents a number of unusual problems for the igneous petrologist. The current exposure represents what is most likely a very small window into a larger magmatic event. Furthermore, the exposed rock represents the roots of the magmatism, most likely at mid- to deep- crustal depths (Elvevold et al., 1994; Tegner et al., 1999), and there are no volcanic rocks preserved from the SIP event. The SIP is composed primarily of mafic plutons of gabbroic composition, together with several large ultramafic bodies. Secondary (but common) magmatism is alkaline in character, and crustal melts such as granites form a relatively insignificant part of the magmatic province.

Understanding the possible settings in which the SIP could have been emplaced requires the identification of analogues in the rock record for such varied magmatism. A variety of igneous provinces are summarised in Table 4.3. As exposure of the SIP is entirely intrusive in character, modern igneous occurrences comprising extensive extrusive volcanism are not ideal analogues for the SIP, while older partially eroded igneous provinces must also be considered.

From Table 4.3, it can be observed that the combination of magmas present in the SIP regularly occurs in a few tectonic settings. Plumes, island arcs and continental subduction zones are not associated with alkaline magmatism in general, although the Canary Islands are unusual in that they do host carbonatites and other alkaline magmas. The variety of magmas occurring in the SIP are regularly associated with magmatism in an extensional setting, be it back-arc rifting or intra-continental rifting.

The Gardar Province in Greenland is an example of an igneous province in an extensional setting. It is of a similar size to the SIP, and comprises numerous plutons of different compositions (Upton et al., 2003). The Gardar represents a rifting event at shallow crustal levels, and the bulk of the magmatism is present as subaerial lavas. However, the spatial density of the intrusions is much lower than that in the SIP and compositionally the lavas also differ considerably from the SIP mafic magmas, in that a clear trend from transitional basalt to trachyte and phonolite is present in the Gardar Province (Upton et al., 2003).

Table 4.3- Selected examples of igneous occurrences

Intrusive Province	Location	Tectonic Setting	Area	Intrusive Composition	Extrusive composition
Iceland	Iceland	Plume	300 km x 200 km	None	Basalt, with minor rhyolitic and alkali basalt
Canary islands	West of East Africa	Plume	30 km x 30 km	None	1) Basalt, trachytes and phonolites, 2) Basanites, nephelinites and minor carbonatite
Mariana Trough	Western Pacific Ocean	Back-arc rifting	1000 km x 100km	None	Calc-alkaline basalt and rhyolite
Gardar	Greenland	Intra-continental Rift	120 km x 300 km	Dolerites, syenites and minor carbonatites and lamprophyres	Basalts, trachytes and phonolites
Trans-Pecos	New Mexico- Texas	Intra-continental Rift	600 km x 200 km	Gabbro, syenites and nepheline syenites	Hawaiites, phonolites and trachytes
Gregory Rift	East Africa	Intra-continental Rift	2000 km x 500 km	Carbonatites and nepheline syenites	Basalts, trachytes and phonolites
Oslo Graben	Norway	Intra-continental Rift	200 km x 50 km	Monzonites, granites, syenites and minor carbonatite	Basalt, trachytes and phonolites
Midcontinental Rift System	North America	Intra-continental Rift	2000 km x 200 km	Differentiated gabbro, nepheline syenite, carbonatite	Basalts and rhyolites
Caledonian Gabbros	Northeast Scotland	Continental Subduction	100 km x 100 km	Differentiated gabbro, peridotite, dunite and norite	None
Volcanic provinces of Italy	Italy	Continental Subduction	500 km x 100 km	None	Potassic alkali basalt, andesites, rhyolites
Andean Coast	South America	Oceanic Subduction	10000 km x 500 km	Gabbro and granite	Basalts and rhyolites
Japan	Japan	Island arc	600 km x 200 km	Minor gabbro and granite	Basalt and rhyolite

Information from Faure, 2001

Carbonatites and alkaline magmas are now generally accepted on isotopic grounds as magmas produced in the mantle by extremely small degrees of partial melting, probably of a metasomatised or CO₂-enriched material (Bell & Blenkinsop, 1987; Tilton & Bell, 1994). As such, alkaline magmas can be produced in a wide variety of tectonic settings, and appear to have become more common with time, as lower heat flow in the mantle favours the low degrees of partial melting required (Blichert-Toft & Albarede, 1997). Although carbonatites and nepheline syenites can be found amongst ocean island basalts (e.g. Canary Islands) and subduction-related granitoids (e.g. Lake Superior Province, Canada), the most common occurrence of such magmas is in rift settings (such as the East African Rift System).

Within the restricted spatial extent of the SIP exposure, alkaline magmatism is relatively abundant. It is also important to note that the magmatic rocks of the SIP are emplaced into continental crust. The mafic plutons of the SIP are variable in composition, comprising both calc-alkaline and tholeiitic gabbro composition. In general, such a mix of mafic magmatism and voluminous alkaline magmatism is accepted as an empirical indicator of extensional magmatism (e.g. Bailey, 1974; Bell & Blenkinsop, 1989; Burke et al., 2003). Magmatism in an extensional stress regime is thus considered the most likely interpretation of the available data for the SIP. However, this extension could have occurred in a variety of settings, and it is impossible to constrain the setting of the SIP to intracontinental rifting, back-arc extension, or an extensional phase of transtension on the basis of the currently available data.

4.3.4. The palaeogeographic significance of the age of the SIP

The SIP has always been a vital part of the tectonic models postulated for the Kalak Nappes and the northern part of the Norwegian Caledonides (e.g. Robins & Gardner, 1974; Ramsay et al, 1985; Daly et al., 1991). The linking of magmatism in the nappes with the evolution of the Caledonides rests on two assumptions: 1) that the Kalak Nappes represent part of the Baltica craton, and 2) that the magmatism in the SIP was related to orogenic deformation. On the basis of these two assumptions, orogenic events within the Kalak Nappes have been extended to cover the whole of the northern Caledonides (e.g. Daly et al., 1991; Elvevold et al., 1994).

Table 4.4. Summary of age dates from this study

INTRUSION	AGE (Ma)	AGE SPREAD (Ma)
Hasvik Gabbro	562 ± 6	556 – 568
Breivikbotn Quartz Diorite	571 ± 4	567 – 575
Storelv Gabbro	569 ± 5	564 – 574
Storelv Granodiorite	563.3 ± 2.1	561 – 565
Øksfjord Gabbro	565.2 ± 9.2	556 – 574
Øksfjord Monzonite	565 ± 5	560 – 570
Øksfjord Monzodiorite	566 ± 4	562 – 570
Øksfjord Norite	569 ± 9	560 – 578
Øksfjord Orthopyroxenite	566 ± 1	565 – 567
Øksfjord Granite	532 ± 31	501 – 563
Breivikbotn Shonkinite	574 ± 5	569 – 579
Breivikbotn Syenite Gneiss	570 ± 2	568 – 572
Breivikbotn Alkaline Gneiss	579 ± 14	565 – 593

The ages determined in this study are summarised in Table 4.4. All the ages obtained in this study fall within a narrow time range, from 555 Ma to 579 Ma, if the limits of uncertainty are used (Fig. 4.5). However, if the minimum spread of overlapping ages is used, the bulk of SIP magmatism could have been emplaced between 563 Ma and 567 Ma without exceeding the error bounds on any age determination. This is significantly shorter than previous estimates (e.g., Sturt, Pringle & Roberts, 1975; Sturt, Pringle & Ramsay, 1978; Daly et al., 1991), and clarifies a number of problems with the current tectonic framework for the region.

As two plutons emplaced at nearly the same time, the Storelv and Hasvik Gabbros illustrate these problems. The Storelv Gabbro is considered to show good structural evidence for syn-kinematic emplacement (Sturt & Taylor, 1971), whereas the Hasvik Gabbro has provided evidence for emplacement in a non-compressional strain regime (Daly et al., 1991). It is unlikely that the Storelv Gabbro was emplaced during constrictional strain (Sturt & Taylor, 1971) at the same time as other gabbros on the island of Sørøy were being intruded without undergoing deformation. Indeed, most outcrops of the Storelv Gabbro reveal well-preserved igneous textures (Sturt & Taylor, 1971; this study), with only small areas showing intense deformation. The discrepancy between the reported deformation in the different gabbros and the coeval age of intrusion for all gabbros would tend to contradict the idea of several extensional deformation events affecting the Kalak Nappes during the emplacement of the SIP.

In general, the deformation recorded in rocks of the SIP can be considered extremely heterogeneous. Undeformed rocks and heavily foliated and metamorphosed rocks are

all the same age. Some of the deformation is likely to be related to the Scandian Phase of the Caledonian Orogeny (420 Ma). If the barometric calculations of 6-8 kbar for the Hasvik Gabbro (Reginiussen, 1996) are correct, then the magma was emplaced into the middle crust under ductile conditions.

Currently, the Finnmarkian Orogeny is defined as the compressive deformational event responsible for the emplacement of the SIP plutons as syn-deformational intrusive bodies (Ramsay et al., 1985). However, this study presents problems for the Finnmarkian concept. As the zircons in the SIP record no deformation between 570 Ma and 420 Ma, the existence of the Finnmarkian Orogeny must be considered questionable. However, what is clear from this study is that even if the Finnmarkian and the preceding Porsanger Orogeny (Daly et al., 1991) did occur, they are not Baltican orogenies, i.e. they did not occur on the margins of Baltica. Thus, the significance generally accorded such events in the Caledonian literature (e.g. Roberts, 2003) must be reassessed.

Chapter 5: A geochemical and isotopic survey of the igneous rocks of the Seiland Igneous Province

5.1. Introduction

In the absence of classical geological features such as contemporaneous sedimentation and distinctive structural controls, comprehensive geochemistry is vitally important in unravelling the geological history of igneous rocks. Such an approach uses both elemental and isotopic chemistry to constrain the possible tectonic environments in which an igneous body or bodies was emplaced, and has been used with some success in a variety of tectonic settings (e.g. Davidson et al., 2005). The Seiland Igneous Province (SIP) is a prime target for such investigation, consisting as it does of a set of varied igneous intrusions confined to a remnant terrane incorporated into the Caledonide orogenic belt. As such, the lack of associated volcanic and sedimentary deposits and the limited volume of magmatism currently preserved across the SIP hampers the identification of the tectonic setting in which the magmatism occurs by traditional methods. In this chapter trace element and isotopic data are presented that will illuminate some issues related to the origin and emplacement of the SIP.

5.1.1. Geochemical investigations of tectonic setting and mantle provenance

Trace element ratios and isotopic signatures have long been used to differentiate magmatic provinces into distinct groups, and numerous tectonic discrimination approaches are presented in Rollinson (1993) and White (2001). These techniques depend on the assumption that distinctive emplacement processes such as contamination, fractional crystallisation and magma mixing produce distinctive geochemical characteristics that are retained in the magmatic products of such events. Such techniques have proved extremely useful in studies of magmatic activity

associated with the oceanic crust (e.g. Halliday et al., 1995; Allegre et al., 1995), and have often been applied to continental settings (e.g. Texeira et al., 2002). The use and misuse of such techniques in identifying both the tectonic setting in which magmatism occurs and in identifying the mantle source for such magmatism has been reviewed by Davidson et al. (2005) and O'Hara and Herzberg (2002). In particular, both contributions emphasise that any magma emplaced into the crust or erupted onto the surface has a long and complicated igneous history, and that it is extremely simplistic to consider such magmatic products to be even remotely representative of the primary magmas produced during partial melting at the source. Magmas undergo both contamination and equilibration with the surrounding rock during ascent, and phase equilibria considerations argue against most surface rocks being primary mantle melts, even if the rocks do preserve mantle-derived isotopic ratios.

Despite these caveats, geochemical data are a valuable tool in investigating the relationship between different igneous rocks. Halliday et al. (1995) and Allegre et al. (1995) showed that the limited REE and trace element variation amongst ocean island basalts indicated a mineralogically homogenous source for such basalts despite the well-known isotopic variation amongst such rocks, whereas Texeira et al. (2002), Kent et al. (2002), and Halama et al. (2004) have used detailed geochemistry to investigate the contamination of rift-related dykes and continental flood basalts. Detailed geochemical and isotopic work is also a feature of many studies of layered mafic intrusions. Such data are available from the Bushveld Complex in South Africa (Maier & Barnes, 1998), the Skaergaard Intrusion in Greenland (Halama, 2002), and the Bjerkreim-Sokndal Intrusion in Norway (Charlier et al., 2005). Most of these studies focus on magmatic processes in the intrusions, but these studies represent

relatively well-understood examples of magmatic events and provide a valuable source for comparative data for provenance studies.

5.1.2. Previous ideas on the origins of the Seiland Igneous Province

The SIP consists of contemporaneous mafic, ultramafic, intermediate, granitic and alkaline intrusions emplaced into a 60 km by 90 km area between 560 and 570 Ma. These intrusions are constrained to a single nappe within the Kalak Nappe complex of Northern Norway. This nappe complex has been generally assumed to be a para-autochthonous terrane within the 420 Ma. Norwegian Caledonides (e.g. Ramsay et al., 1985), but more recent work has indicated that the terrane may be exotic and allochthonous (Corfu et al., 2006). The largest component of the SIP consists of numerous mafic plutons, commonly layered, which comprise at least 60% of the province. Large ultramafic complexes comprise a further 25% of the complex, with intermediate rock types such as monzonite and diorite comprising a further 10% of the complex. Alkaline intrusions run throughout the province (5%), but granitic rocks are restricted to a few small, insignificant bodies on Øksfjord and Sørøy.

In the only general reviews of the igneous activity within the SIP, Robins and Gardner (1974, 1975) postulated that the SIP represented the remains of an orogenic Andean-style magmatic episode. This view was based on the available age dating at the time, which indicated a long emplacement period (± 100 Ma), and the authors' subdivision of the gabbroic plutons into several distinct subsets on mineralogical grounds. In this model, different types of intrusions were emplaced at intervals into the underlying Kalak Nappe during episodic subduction along a continental margin.

This view was largely superseded by the first Sm-Nd dating to emerge from the area during the 1980s and 1990s, which implied that the magmatism in the area had started significantly earlier than previously thought (ca. 800 Ma), and had continued over a period of 300 Ma (Mørk & Stabel, 1990; Daly et al., 1991; Cadow, 1993). These ages were used to argue for a scheme of protracted rifting in the area (e.g. Elvevold et al., 1994; Reginiussen, 1996), in which the gabbroic rocks of the SIP were emplaced in pulses over a long period of continental rifting. This age scheme has since been superseded by detailed U-Pb TIMS data (Chapter 2; Roberts et al., 2006) implying a short (10 m.y.) period of intrusion at 570 Ma.

In a review of the mafic rocks comprising the Seiland Igneous Province, Robins & Gardner (1974, 1975) divided the gabbroic plutons present into three categories: those they considered to have crystallised from a tholeiitic parental melt, those they considered to have crystallised from an alkali basalt parent, and those they defined as “syeno-gabbros”, with monzonites and syenites interlayered with tholeiitic gabbro. This classification scheme has not been superseded (e.g. Robins, 1996).

Problems in the classification scheme arise from the allocation of parental magma types to the different gabbros, with an assumption that different parental magma compositions were erupted at different times. This scheme has now been invalidated by the new geochronology. Furthermore, Robins (1982) notes that the Rognsund clinopyroxene gabbro has undergone extensive fractionation prior to emplacement, and so the original parental magma composition, assumed to be of an alkaline olivine basalt composition by Robins & Gardner (1975), has been extensively modified during its ascent into the crust. It is thus unclear whether one or more magma sources

are involved in the SIP, or whether the variation amongst the gabbroic plutons can be related to different rates of ascent, fractionation and crustal contamination.

5.1.3. Previous isotopic and geochemical work on the SIP

Currently available isotopic results for the Province are limited in extent, and detailed REE and trace element chemistry for any of the SIP intrusions has not yet been collected. Published Rb-Sr data are either incompletely reported (Sturt et al., 1976, 1978) or are extremely limited (Krogh & Elvevold, 1991), whereas the use of Sm-Nd data for provenance studies has been hampered by the lack of accurate dating (e.g. Daly et al., 1991; Cadow, 1993; Tegner et al., 1999). Furthermore, the isotopic analyses have been confined to the mafic rocks of the province, and so cannot address the relationships between the mafic, intermediate and alkaline rocks in the province.

Relatively detailed mineralogical studies have been conducted on the Lille Kufjord (Aitcheson & Forrest, 1994) and the Hasvik Intrusion (Robins & Gardner, 1974; Tegner et al., 1999). These studies have focused on the magmatic evolution of specific plutons, and both studies have reported large degrees of crustal contamination. In the only detailed isotopic study of a SIP intrusion, Tegner et al (1999) report Sm-Nd and Rb-Sr data from the Hasvik Layered Intrusion. This primarily gabbroic intrusion displays increasing $^{87}\text{Sr}/^{86}\text{Sr}$ and decreasing $^{147}\text{Nd}/^{144}\text{Nd}$ ratios with increasing height in the chamber, and evolves from mantle isotopic values ($^{87}\text{Sr}/^{86}\text{Sr} = 0.7038$, $\epsilon_{\text{Nd}} = +4.76$) to values indicative of high degrees of crustal contamination. This isotopic behaviour can be correlated with a decreasing whole rock Mg# (0.7- 0.3) and a decreasing An component in plagioclase (0.72-0.52). Modelling undertaken by Tegner et al. (1999) assumed a quartzofeldspathic country

rock contaminant, and produced an estimate of 21% crustal contamination for the Hasvik Layered Intrusion.

Profiles of major and trace element chemistry are available for two of the plutons in the SIP, though the trace elements are limited in both cases. The Rognsund Gabbro on Seiland, considered by Robins & Gardner (1974) to be a clinopyroxene gabbro derived from an alkali olivine basalt parent, has been investigated by Robins (1982), and the Hasvik Gabbro, considered to be the product of the fractionation of a quartz-normative tholeiitic basalt parent, has been investigated by Tegner et al. (1999). The two plutons show very different fractionation trends from one another. The Hasvik Gabbro shows the development of a thin basal zone containing numerous country rock xenoliths, but the bulk of the intrusion consists of a layered sequence evolving from a primitive Lower Zone through a Main Zone to an Upper Zone. Assimilation-fractional crystallisation processes drive the evolution of this sequence, and the Upper Zone is held to comprise nearly 20% assimilated crustal material, reflected in its silica-rich nature (Tegner et al., 1999). The Rognsund Gabbro consists of a thick (150 m) xenolithic Contaminated Zone, and a fractionated Layered Series.

5.1.4. Outline of the current study

This study reports new major and trace element data for nine mafic intrusions in the SIP, three ultramafic intrusions, five intermediate or granitic bodies, and two nepheline syenites from the SIP. This study makes use of previously published Sm-Nd data as well as new Sm-Nd data, and also reports new Lu-Hf isotopic analyses obtained from zircon separates collected in the course of the U-Pb TIMS dating project. These data are used to investigate the interrelationships between the different

plutons and the associated rocks, and are also compared with data from a variety of magmatic events from a variety of tectonic settings, in order to constrain the possible emplacement environment for the SIP.

5.2. Sampling methodology

Samples for geochemistry were selected from a variety of localities across the SIP for this study (Figure 5.1). Relatively unweathered samples were preferred, although some rocks do show limited development of alteration textures along grain boundaries. Although as many igneous bodies as possible were sampled, there are areas that are not covered by the sample set, notably the islands of Seiland and Stjernøy. The sample set is not comprehensive but has been selected to maximise the variety of rocks sampled. This can be compared to the data sets reported by Robins (1982) and Tegner et al. (1999), which comprise profiles through gabbroic plutons with clearly defined fractionation trends.

On Sørøy, samples were taken from the Hasvik Intrusion (RJR-02-30B), the Breivikbotn Intrusion (RJR-03-114B and the Breivikbotn Nepheline Syenite Complex (RJR-03-115 and 116). The remainder of the samples come from the Øksfjord peninsula, where the field relationships between different intrusions are obscure. As such, the Øksfjord samples were selected to represent as broad a range of rock types as possible, ranging from orthopyroxenites and wehrlites to norites and gabbros, along with several samples of monzonites and diorites. Of special interest is the presence of two samples from the cross-cutting Tappeluft Complex at the southern end of the Øksfjord peninsula, monzonite RJR-02-8 and gabbro RJR-03-109. Most of the

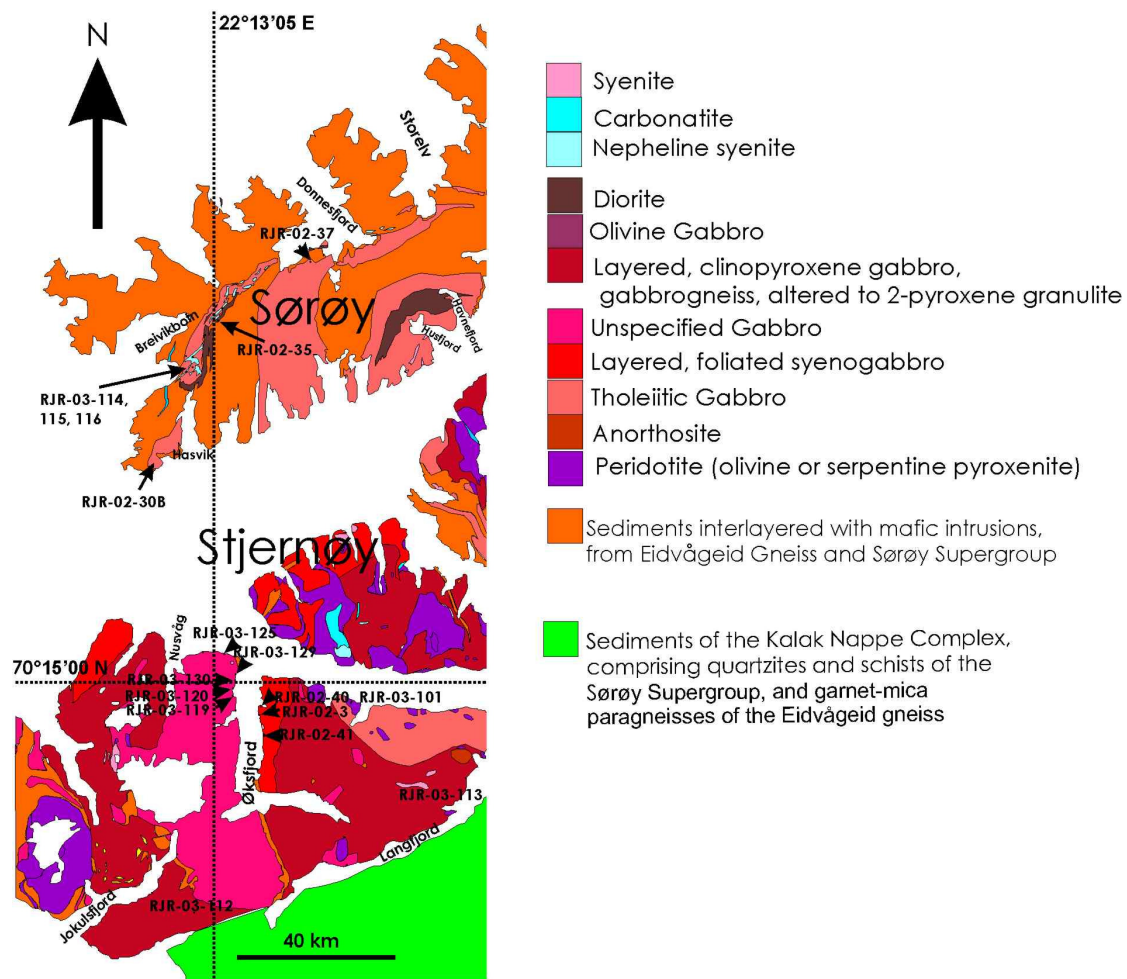


Figure 5.1- Location of sampling points for geochemical investigation.

gabbroic rocks are currently classified as “syenogabbros” under the scheme of Robins & Gardner (1974), though the Hasvik Gabbro is classified as a “tholeiitic” gabbro and the Breivikbotn Gabbro is currently unclassified.

Of the nineteen rocks analysed for their geochemistry, ten have been dated directly by U-Pb TIMS to the period 560- 570 Ma (Chapter 4). The other rocks are taken from intrusions in close proximity to the dated intrusions, and are considered to be coeval with the dated rocks on petrological grounds. Amongst the mafic rocks, RJR-03-109, RJR-03-113B, RJR-03-114B, RJR-03-120A, RJR-03-125, and RJR-03-130 all display

little evidence of strain and have characteristic spinel-pyroxene symplectites considered to represent a decompression reaction between olivine and plagioclase, a texture which is also found in most of the dated mafic rocks. This texture is taken as an indication that all the rocks have experienced a similar post-emplacement metamorphic history, involving emplacement at 560 Ma and deformation at upper amphibolite facies at 420 Ma, and can thus be considered no younger than the dated rocks.

Though the dating of ultramafic rocks is always difficult, orthopyroxenite RJR-03-129D was dated, and fragments of zircon were recovered from RJR-03-108 which indicated a SIP age in the region of 560-570 Ma (Chapter 4). Amongst the intermediate rock types, only RJR-03-119B was not directly dated, but this diorite is emplaced into a mafic gabbro considered to be of SIP age (RJR-03-120A) and has been metamorphosed along with the gabbro. Nepheline syenite RJR-03-115 is nearly identical petrographically to RJR-03-116 and was emplaced in close proximity and in a similar manner (Chapter 4), but proved to be barren of zircon for dating purposes. It is however considered to be of the same age as RJR-03-116.

The supporting isotopic data set does not necessarily correlate with the geochemical data set, which is primarily based on the ID-TIMS zircon sampling programme. As such, the Lu-Hf data cover rocks not included in the geochemical data set and vice versa. The Sm-Nd data set is extremely small and was selected to solve specific isotopic problems and confirm observations made on other portion of the data, such as the relationship between the two nepheline syenites (which show significant trace element difference and of which only one has been dated). This limited Sm-Nd data

set has been supplemented with data taken from the literature and recalculated to SIP ages.

5.3. Results

5.3.1. Geochemical Analysis

Tables 5.1 and 5.2 show the major and trace element data from a selection of rocks from the SIP. Table 5.3 shows the CIPW norms for each rock. In general, the CIPW norm corresponds well with the observed mineralogy in these rocks. In some cases, the presence of symplectites within a rock and uncertainty as to the true $\text{Fe}^{2+}/\text{Fe}^{3+}$ ratio in the rocks means that the norm is a little different to the mode given above, but the two estimates of mineralogical composition are generally very close.

Bivariate Harker plots of SiO_2 versus other major elements are shown in Figure 5.2 to show the general variability amongst the SIP gabbros. Linear trends are unlikely to be present in Figure 5.2 unless every pluton is derived from identical magmas under identical conditions. However, anomalous rocks and the general spread of values can be seen in such diagrams. From the plots in Figure 5.2, it is possible to make some broad comments about the plutons of the SIP.

Table 5.1. Major element data for rocks from the SIP (sampling sites on Figure 2.1)

	RJR-02-30B <i>Hasvik</i> <i>Gabbro Upper</i> <i>Zone</i> Ne- Normative Olivine Gabbro	RJR-03-114 <i>Breivikbotn</i> <i>Gabbro</i> Olivine Norite	RJR-03-102 <i>Øksfjord</i> <i>Gabbro</i> Ne- Normative Olivine Gabbro	RJR-03-125 <i>Øksfjord</i> <i>Gabbro</i> Gabbronorite	RJR-03-112 <i>Øksfjord</i> <i>Gabbro</i> Ne- Normative Olivine Gabbro	RJR-03-120A <i>Øksfjord</i> <i>Gabbro</i> Olivine Gabbronorite	RJR-03-129C <i>Øksfjord</i> <i>Gabbro</i> Olivine Gabbronorite	RJR-03-130 <i>Øksfjord</i> <i>Gabbro</i> Ne- Normative Olivine Gabbro	RJR-03-109 <i>Øksfjord</i> <i>Gabbro</i> Ne- Normative Olivine Gabbro	RJR-03-113B <i>Langfjorden</i> <i>Gabbro</i> Ne- Normative Olivine Gabbro
SiO ₂	48.13	47.69	44.77	47.34	47.15	49.92	48.96	46.46	41.41	46.46
Al ₂ O ₃	16.95	13.88	16.81	14.92	14.96	17.04	14.63	16.12	18.51	16.27
Fe ₂ O ₃	1.21	1.28	1.39	1.52	1.41	1.19	1.51	1.51	1.44	0.7
FeO	9.78	10.37	11.29	12.29	11.38	9.65	12.22	12.25	11.66	5.68
MnO	0.1663	0.2077	0.2499	0.2163	0.2024	0.1737	0.2249	0.1874	0.1651	0.1206
MgO	6.05	10.96	4.77	6.05	5.65	5.96	4.63	5.11	6.31	12.8
CaO	10.43	9.88	10	10.7	10.2	9.93	9.03	9.5	10.72	15.03
Na ₂ O	3.39	1.35	4.08	2.4	3.17	3.26	3.82	3.56	3.57	1.23
K ₂ O	0.63	0.67	1.46	0.69	1.01	0.4	0.63	1.32	0.85	0.08
TiO ₂	1.9654	2.3388	3.1244	3.4869	3.4438	2.436	3.3619	2.8749	4.2074	0.4659
P ₂ O ₅	0.25	0.5	1.72	0.54	0.6	0.4	0.59	0.47	0.55	0.02
Cr ₂ O ₃	0.0487	0.0887	0.0164	0.0196	0.0231	0.0295	0.0136	0.0419	0.0124	0.0919
NiO	<u>0.0201</u>	<u>0.0355</u>	<u>0.0056</u>	<u>0.0076</u>	<u>0.0132</u>	<u>0.0051</u>	<u>0.0035</u>	<u>0.0054</u>	<u>0.0055</u>	<u>0.0306</u>
TOTAL	99.02	99.25	99.7	100.18	99.21	100.42	99.62	99.4	99.41	98.98
Mg#	52.43	65.32	42.95	46.73	46.94	52.39	40.30	42.64	49.09	80.06

Table 5.1. continued

	RJR-03-108A <i>Tappeluft Ultramafic Complex</i> Ne-Normative Wehrlite	RJR-03-129D <i>Øksfjord Gabbro</i> Olivine Orthopyroxenite	RJR-03-119B <i>Øksfjord Gabbro</i> Diorite	RJR-03-129A <i>Øksfjord Gabbro</i> Granite	RJR-02-41C <i>Øksfjord Gabbro</i> Monzodiorite	RJR-02-40B <i>Øksfjord Gabbro</i> Monzonite	RJR-02-08 <i>Tappeluft Ultramafic Complex</i> Monzonite	RJR-03-115 <i>Breivikbotn Alkaline Complex</i> Nepheline Monzodiorite	RJR-03-116 <i>Breivikbotn Alkaline Complex</i> Nepheline Monzodiorite
SiO ₂	40.21	53.59	60.57	78.73	62.25	63.28	63.78	60.67	59.5
Al ₂ O ₃	4.75	4.67	23.42	11.27	18.24	18.27	19.54	23.43	23.44
Fe ₂ O ₃	1.97	1.34	0.15	0.18	0.41	0.34	0.22	0.21	0.21
FeO	15.98	10.86	1.23	1.46	3.32	2.74	1.8	1.69	1.7
MnO	0.2701	0.2009	0.036	0.0275	0.1246	0.124	0.0263	0.0978	0.0575
MgO	28.44	24.74	0.69	0.33	0.55	0.49	0.3	0.05	0.06
CaO	6.53	0.87	5.52	1.29	2.87	2.15	1.59	0.57	0.49
Na ₂ O	0.43	1.17	7.25	3.04	6.02	5.7	5.07	9.73	9.89
K ₂ O	0.04	1.54	1.08	2.88	5.27	6.53	7.44	3.68	4.63
TiO ₂	0.6577	0.2718	0.2309	0.7424	0.3977	0.4007	0.4162	0.0705	0.0947
P ₂ O ₅	0.03	0.13	0.05	0.07	0.18	0.14	0.07	0.02	0.03
Cr ₂ O ₃	0.2408	0.076	0.0073	0.0108	0.024	0.0082	0.0042	0.0074	0.0055
NiO	<u>0.108</u>	<u>0.1889</u>	<u>0.0038</u>	<u>0.0006</u>	<u>0.0008</u>	<u>0.0014</u>	<u>0.0006</u>	<u>-0.0004</u>	<u>0.0008</u>
TOTAL	99.67	99.66	100.25	100.02	99.65	100.16	100.26	100.21	100.11
Mg#	76.03	80.24	49.99	28.71	22.79	24.17	22.90	5.01	5.92

Table 5.2. Trace element data for rocks from the SIP

	RJR02-30B Ne- Normative Olivine Gabbro	RJR03-114 Olivine Norite	RJR03-102 Ne- Normative Olivine Gabbro	RJR03-125 Gabbronorite	RJR03-112 Ne- Normative Olivine Gabbro	RJR03-120A Olivine Gabbronorite	RJR03-129C Olivine Gabbronorite	RJR03-130 Ne- Normative Olivine Gabbro	RJR03-109 Ne- Normative Olivine Gabbro	RJR03-113B Ne- Normative Olivine Gabbro
Rb	12.7	15.7	31.2	10.5	26.3	1.6	4.0	24.1	6.3	1.0
Sr	647.1	587.6	1030.3	554.0	536.6	546.3	557.0	538.3	1981.4	421.4
Y	16.5	33.7	39.7	47.4	40.7	26.8	40.9	57.4	23.7	5.7
Zr	104.8	129.1	130.5	336.3	262.6	360.6	374.8	301.2	50.3	9.8
Ba	198.9	424.4	570.1	617.2	345.7	237.7	180.1	292.8	651.2	44.8
La	14.0	32.8	64.9	95.4	34.4	42.6	38.8	41.1	17.0	1.4
Ce	31.6	77.1	134.0	115.8	78.1	90.5	86.5	103.7	41.9	3.0
Pr	4.1	9.8	16.5	15.7	10.4	11.5	11.2	14.3	6.3	0.5
Nd	17.3	39.8	64.4	67.0	43.7	47.0	47.0	62.4	30.5	2.9
Sm	3.8	8.1	11.9	14.3	9.6	9.3	10.0	13.9	6.8	0.9
Eu	1.7	1.9	4.0	3.1	2.6	2.7	2.7	3.6	2.8	0.5
Gd	3.8	7.8	11.5	13.6	9.3	8.8	10.1	13.8	6.3	1.0
Tb	0.5	1.1	1.5	1.8	1.4	1.1	1.4	2.0	0.9	0.2
Dy	3.0	6.0	7.5	9.2	7.4	5.4	7.5	10.8	4.6	1.0
Ho	0.6	1.1	1.3	1.7	1.4	1.0	1.4	2.0	0.8	0.2
Er	1.5	2.9	3.4	4.5	3.4	2.5	3.6	5.0	2.0	0.5
Tm	0.2	0.4	0.5	0.7	0.5	0.4	0.5	0.7	0.3	0.1
Yb	1.3	2.4	2.7	4.0	2.8	2.2	3.0	4.0	1.4	0.4
Lu	0.2	0.4	0.4	0.6	0.4	0.3	0.4	0.5	0.2	0.1
Hf	2.6	3.4	2.6	7.9	6.0	8.1	7.9	7.0	1.5	0.4
Th	1.2	2.5	4.8	1.5	0.7	1.5	1.3	1.2	0.5	0.0
U	0.5	0.9	1.6	0.4	0.7	0.6	0.9	0.6	0.3	0.2
Nb	16.3	36.4	48.8	52.5	58.0	38.8	69.8	48.0	39.3	0.4
Ta	0.9	2.0	2.8	2.8	3.3	2.2	3.8	2.5	2.0	0.0
W	0.3	0.3	0.5	0.3	0.4	0.3	0.5	0.3	0.3	0.2
Pb	2.9	5.3	4.5	5.5	4.3	6.2	4.8	4.1	2.6	2.7

Table 5.2 continued

	RJR03-108A Ne-Normative Wehrlite	RJR03-129D Olivine Orthopyroxenite	RJR03-119B Diorite	RJR03-129A Granite	RJR03-101A Monzodiorite	RJR03-101E Monzonite	RJR02-08 Monzonite	RJR03-115 Nepheline Monzodiorite	RJR03-116 Nepheline Monzodiorite
Rb	0.3	64.8	7.3	85.4	63.1	104.3	71.3	95.8	137.5
Sr	115.1	55.9	2256.6	213.8	309.9	157.6	158.8	90.4	66.3
Y	7.8	16.0	4.5	28.5	13.2	16.0	3.4	44.2	6.4
Zr	20.1	98.1	754.8	927.8	65.0	267.2	16.6	1416.5	40.5
Ba	25.5	190.4	1161.4	651.8	1652.0	950.9	528.3	62.1	95.8
La	4.4	15.0	30.9	38.4	23.1	24.0	15.3	104.9	88.0
Ce	6.1	31.1	37.7	87.7	39.2	48.5	20.1	176.7	124.3
Pr	1.0	3.6	3.1	10.5	4.8	6.2	2.0	16.0	10.0
Nd	5.1	12.4	8.1	39.5	18.9	25.2	7.0	41.7	22.8
Sm	1.4	2.6	1.4	7.8	4.4	5.1	1.4	6.8	2.6
Eu	0.5	0.4	2.8	1.4	3.2	1.9	2.5	1.6	1.1
Gd	1.5	2.6	1.3	7.1	3.6	4.5	1.2	7.6	3.3
Tb	0.2	0.4	0.1	0.9	0.5	0.6	0.1	1.0	0.3
Dy	1.4	2.4	0.6	4.8	2.6	3.2	0.7	6.3	1.2
Ho	0.3	0.5	0.1	0.9	0.5	0.6	0.1	1.4	0.2
Er	0.7	1.5	0.5	2.7	1.2	1.5	0.3	4.5	0.6
Tm	0.1	0.2	0.1	0.4	0.2	0.2	0.0	0.8	0.1
Yb	0.6	1.6	0.6	2.9	1.0	1.5	0.2	5.4	0.6
Lu	0.1	0.3	0.1	0.5	0.2	0.3	0.0	0.8	0.1
Hf	0.7	2.5	10.0	22.1	1.2	5.0	0.4	23.3	0.7
Th	0.1	5.2	0.9	14.2	0.3	0.8	0.3	45.5	5.9
U	0.2	2.1	0.6	3.0	0.4	0.5	0.3	13.3	1.8
Nb	0.8	10.9	18.3	20.0	22.1	28.8	14.5	239.2	39.8
Ta	0.1	0.8	0.9	1.3	1.0	1.3	0.7	21.3	2.5
W	0.2	0.6	0.2	0.4	0.2	0.2	0.2	0.6	0.5
Pb	2.4	5.8	9.1	16.4	9.6	6.0	8.7	11.8	6.8

Table 5.3. CIPW norms for rocks from the SIP

	RJR02-30B	RJR03-114	RJR03-102	RJR03-125	RJR03-112	RJR03-120A	RJR03-129C	RJR03-130	RJR03-109	RJR03-113B
Ne	1.28		7.61		11.73	0.38			4.51	1.11
Q										
Or	3.72	3.96	8.62	2.36	5.02	5.97	4.08	3.72	7.80	0.47
Pl	55.49	41.26	43.72	58.26	40.53	49.73	48.21	53.23	45.91	46.99
Cor										
Cpx	17.28	12.91	12.62	13.16	14.51	19.11	17.95	16.76	16.67	28.60
Opx		28.28		13.52			13.45	6.60		
Ol	15.11	5.26	15.15	5.77	16.25	14.02	6.39	9.34	15.73	19.72
Ap	0.55	1.09	3.76	0.87	1.20	1.31	1.18	1.29	1.03	0.04
Mt	1.75	1.86	2.02	1.73	2.09	2.04	2.20	2.19	2.19	1.01
Il	3.73	4.44	5.93	4.62	7.99	6.54	6.62	6.39	5.46	0.88
Chr	0.07	0.13	0.02	0.04	0.02	0.03	0.03	0.02	0.06	0.13
	98.98	99.19	99.45	100.33	99.34	99.13	100.11	99.54	99.36	98.95
	RJR03-108A	RJR03-129D	RJR03-119B	RJR03-129A	RJR03-101A	RJR03-101E	RJR02-08	RJR03-115	RJR03-116	
Ne	1.68							14.06	20.58	
Q			0.72	46.35			1.03			
Or	0.24	9.10	6.38	17.02	31.14	38.59	43.97	21.75	27.36	
Pl	11.44	12.84	88.40	31.67	58.12	53.21	50.33	59.08	47.92	
Cor			0.41	0.97			0.42	2.45	1.34	
Cpx	17.11	0.42			5.07	4.11				
Opx		59.06	3.54	2.17	1.81	0.37	3.22			
Ol	64.65	15.35			1.71	2.31		2.40	2.34	
Ap	0.07	0.28	0.11	0.15	0.39	0.31	0.15	0.04	0.07	
Mt	2.86	1.94	0.22	0.26	0.59	0.49	0.32	0.30	0.30	
Il	1.25	0.52	0.44	1.41	0.76	0.76	0.79	0.13	0.18	
Chr	0.35	0.11	0.01	0.02	0.03	0.01	0.01	0.01	0.01	
	99.65	99.62	100.23	100.02	99.62	100.16	100.24	100.22	100.10	

- Normative composition is calculated by the CIPW norm. Mineral names: Ne = nepheline, Or = orthoclase, Pl = plagioclase, Cpx = clinopyroxene, Opx = orthopyroxene, Ol = olivine, Ap = apatite, Mt = magnetite, Il = ilmenite, Chr = chromite

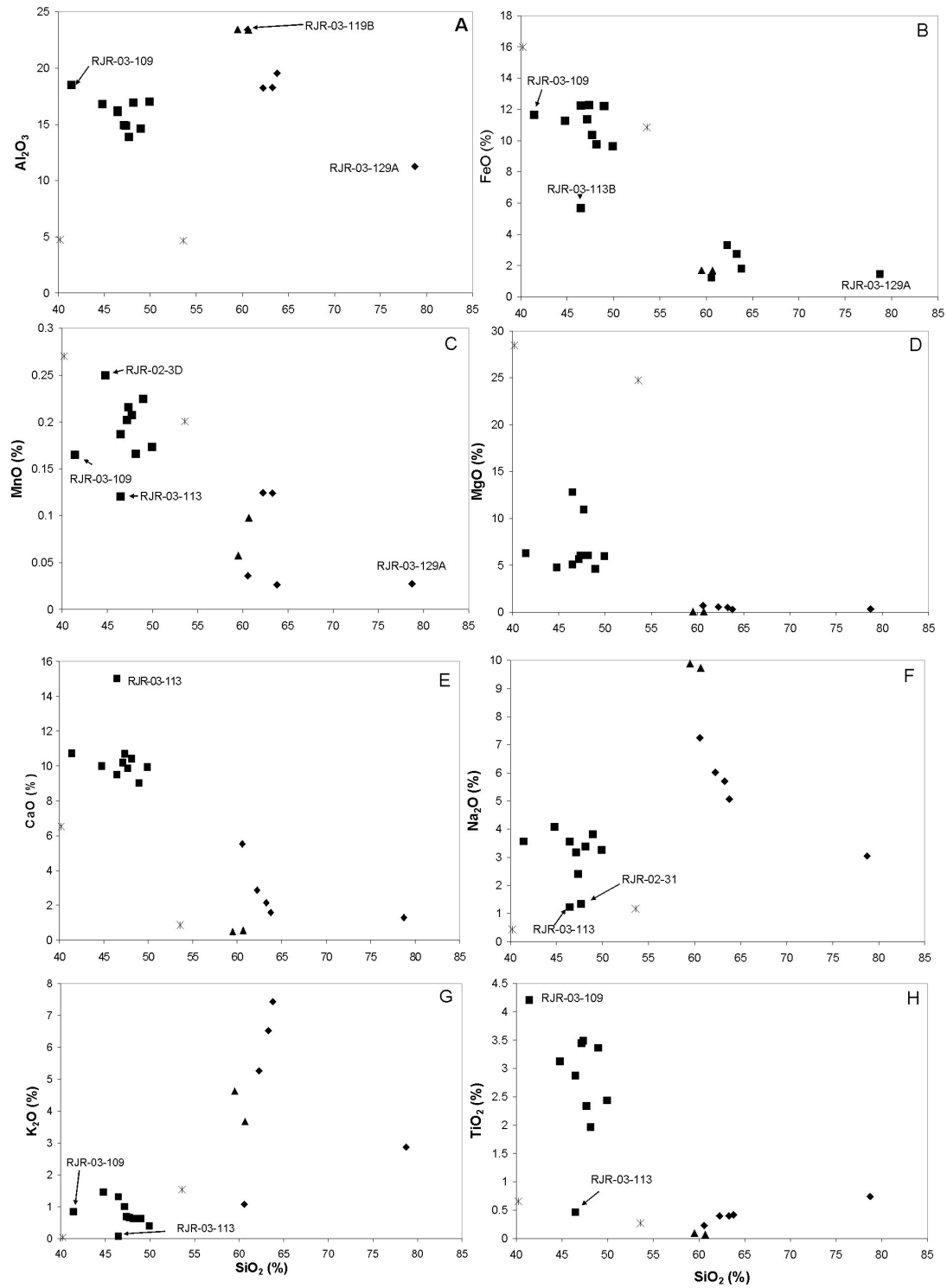


Figure 5.2- Harker plots of data from Table 5.1. Mafic rocks are represented with squares, intermediate and granitic rocks with diamonds, ultramafic rocks with asterisks, and nepheline syenites with triangles.

In general, the mafic rocks range between 44 and 50% SiO₂, between 13 and 17% Al₂O₃, and between 2 and 3.5% TiO₂. One rock, olivine gabbro RJR-03-109 from the Tappeluft Ultramafic Complex on Øksfjord, has the lowest SiO₂ content (41.4 %), and highest Al₂O₃ (18.5%) and TiO₂ (4.2%) relative to the other rocks in the SIP. In contrast, olivine gabbro RJR-03-113 is depleted in many elements compared to the other mafic rocks, but is the most enriched in MgO. The low content of K₂O, P₂O₅ and TiO₂ coupled with a Mg# of 80 indicate that this rock is more primitive than suggested by the normative and modal observations. Reference to the thin section (Chapter 2) shows that a large portion of the rock now consists of orthopyroxene-spinel intergrowths formed by metamorphism of olivine, indicating that the rock originally contained significantly more olivine than is now present. RJR-03-113 is relatively high in SiO₂ and Al₂O₃ compared to the other ultramafic rocks.

Amongst the other rocks investigated, most plot according to their mineralogy. The two nepheline syenites behave similarly in all plots, as they are high in Al and Na but low in most other elements. The granitic rock RJR-03-129A plots separately from the other rocks, as might be expected by a rock dominated by quartz whereas the other intermediate rocks are dominated by feldspars. The three monzonitic rocks plot away from the dioritic rock (RJR-03-119), and the intermediate rocks are clearly separated from the mafic rocks.

The trace element data for the SIP rocks, as presented in Table 5.2, can be examined in various different ways. Figure 5.3 shows a series of bivariate diagrams illustrating the variation between various pairs of trace elements. In some diagrams, such as Figure 5.3A (Rb vs. Sr), a strong mineralogical control on the two elements can be

observed in the separation of rocks into gabbroic (K-feldspar not present) and intermediate (K-feldspar-bearing) rocks.

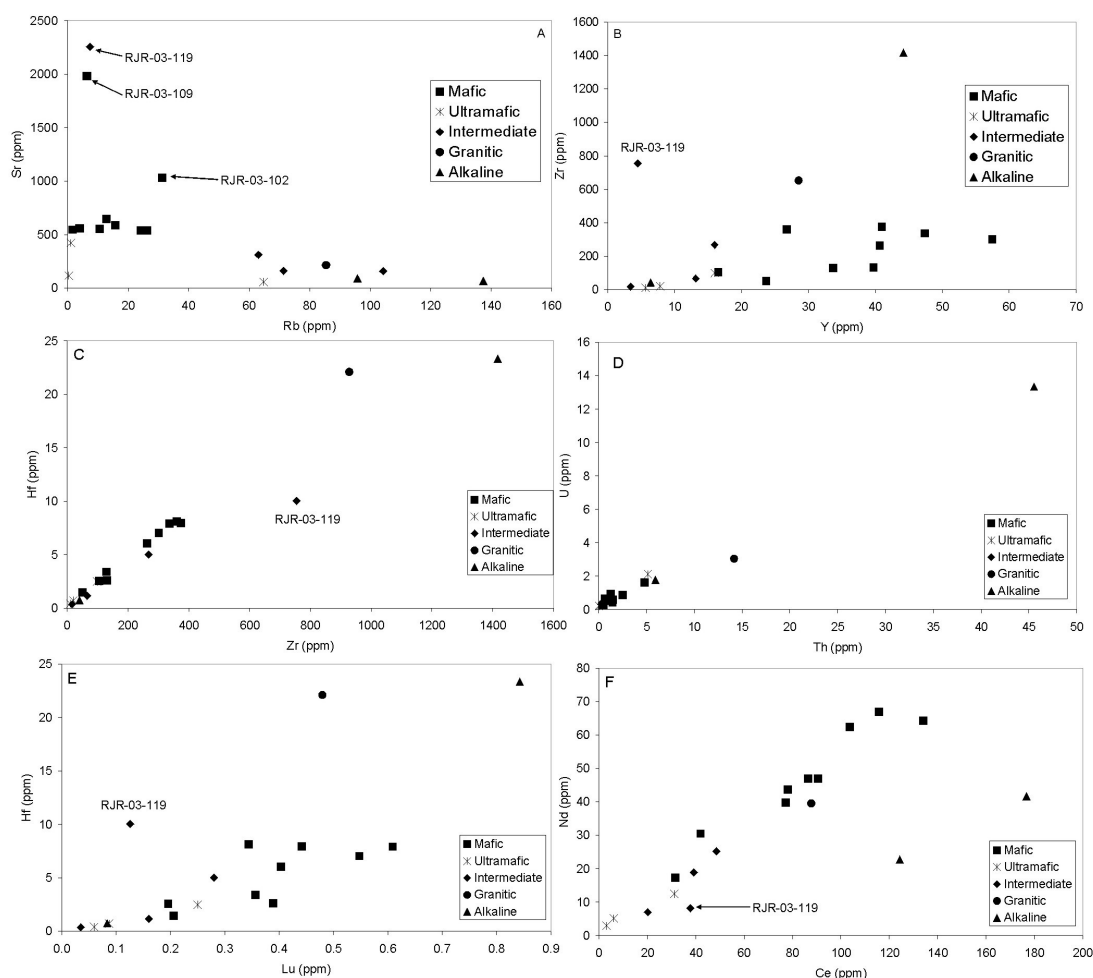


Figure 5.3- Bivariate diagrams for trace elements from SIP rocks. Data from Table 5.2.

In general, most of the igneous rocks in the SIP show similar variation in their trace element contents, regardless of their mineralogical composition. One rock, diorite RJR-03-119, diverges from the other rocks in the data set, and is also clearly different from the granitoid, RJR-03-129A, which is known from U-Pb TIMS investigation to have inherited some components from an older protolith (Chapter 4). A further observation is that the trace element abundances of the two alkaline rocks, nepheline

syenites RJR-03-115 and 116 differ radically, despite their nearly identical mineralogical, chronological and geographical natures.

One set of important trace elements is the rare-earth elements. These elements are generally immobile during weathering and metamorphism. Figures 5.4 – 5.7 show the rare-earth patterns for the different rocks of the SIP, normalised to the chondritic values of Sun & McDonough (1989). Figure 5.4 shows the mafic rocks of the SIP, to which the other rocks in the province are compared. The mafic rocks show a negative slope from La to Lu, with only minor Eu anomalies. Furthermore, the rocks show a limited variation in their REE contents, and are all enriched relative to profiles reported from other continental mafic intrusions (see discussion below).

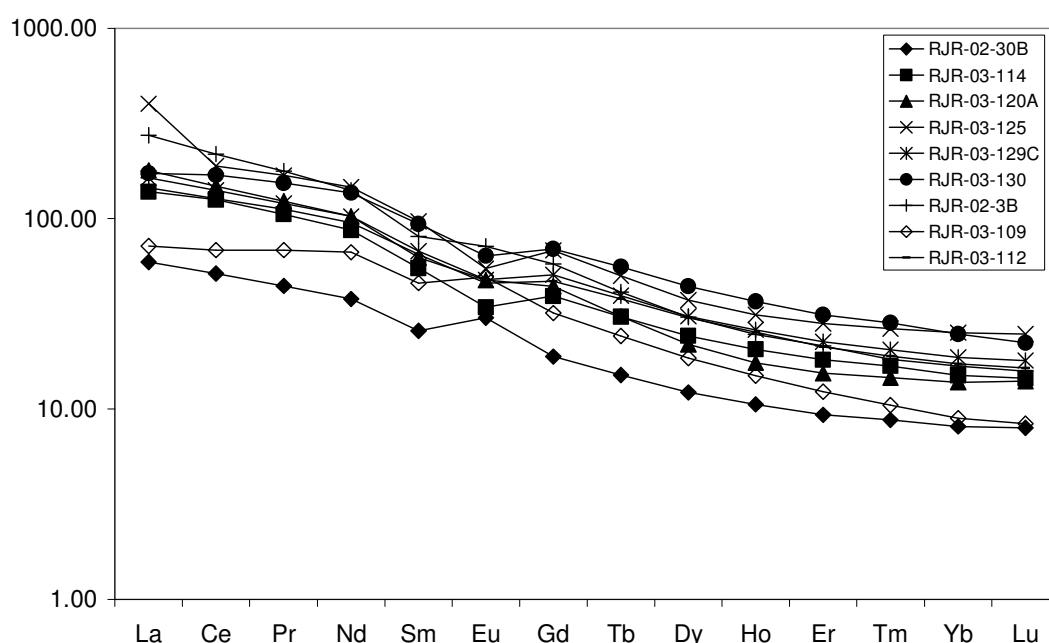


Figure 5.4- Rare earth element plot for the mafic rocks of the SIP. Data from Table 5.2, normalised to the chondritic values of Sun & McDonough (1989)

Figure 5.5 shows the ultramafic rocks of the SIP in comparison to the mafic rocks. Also included in this plot is RJR-03-113B, the olivine gabbro identified in the major element plots as significantly different from most of the other mafic rocks. It can be

seen that all three rocks are less abundant in the rare earths than the mafic rocks. Furthermore, it can be observed that orthopyroxenite RJR-03-129D is closest to the mafic rocks in pattern, and is relatively light REE enriched profile but a prominent negative Eu anomaly.

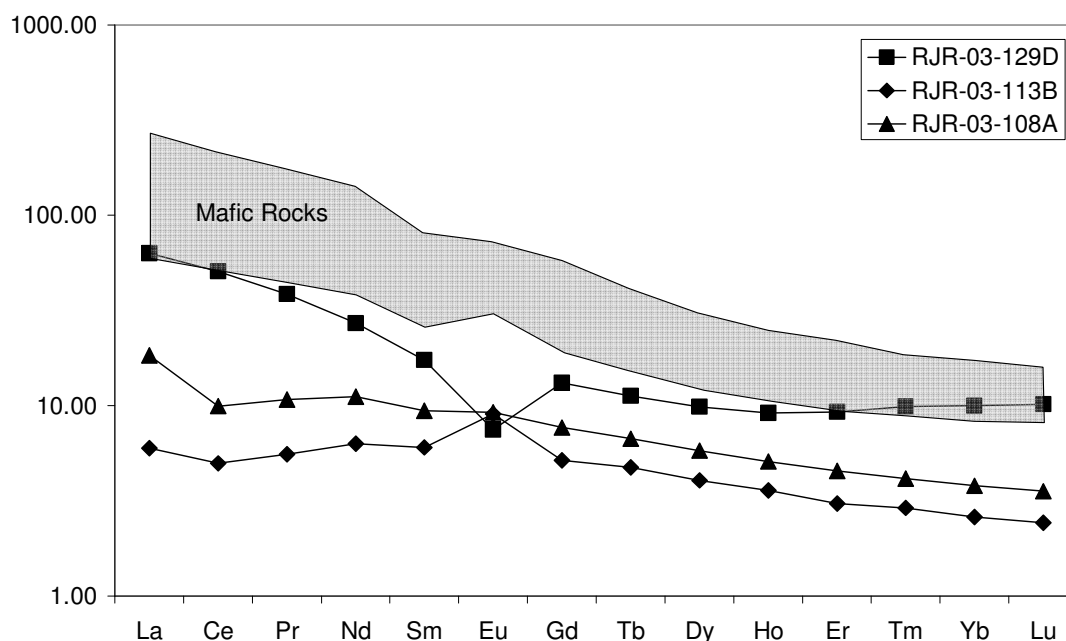


Figure 5.5- REE profiles for ultramafic rocks from the SIP. Data from Table 5.2, normalised to the chondritic values of Sun & Mcdonough (1989). REE profile for the mafic rocks taken from Figure 5.4.

Figure 5.6 displays the REE profiles for the intermediate and granitic rocks of the SIP. Granite RJR-03-129A and monzonites RJR-03-101A and RJR-03-101E are similar in profile to the mafic rocks. Monzonite RJR-02-08 and diorite RJR-03-119 show significantly different REE profiles, with much larger positive Eu anomalies and a steeper slope between La and Lu than is present in the mafic rocks. The positive Eu anomaly in these rocks is unusual and needs to be addressed.

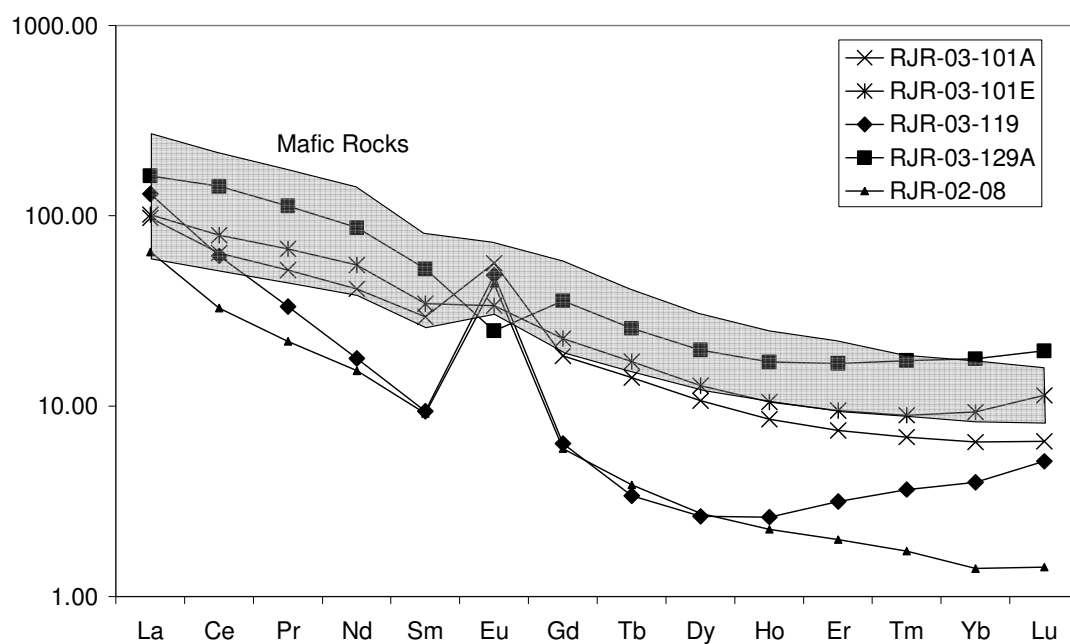


Figure 5.6- REE profiles for intermediate and granitic rocks from the SIP. Data from Table 5.2, normalised to chondritic values from Sun and McDonough (1989). Mafic rock profile taken from figure 5.4.

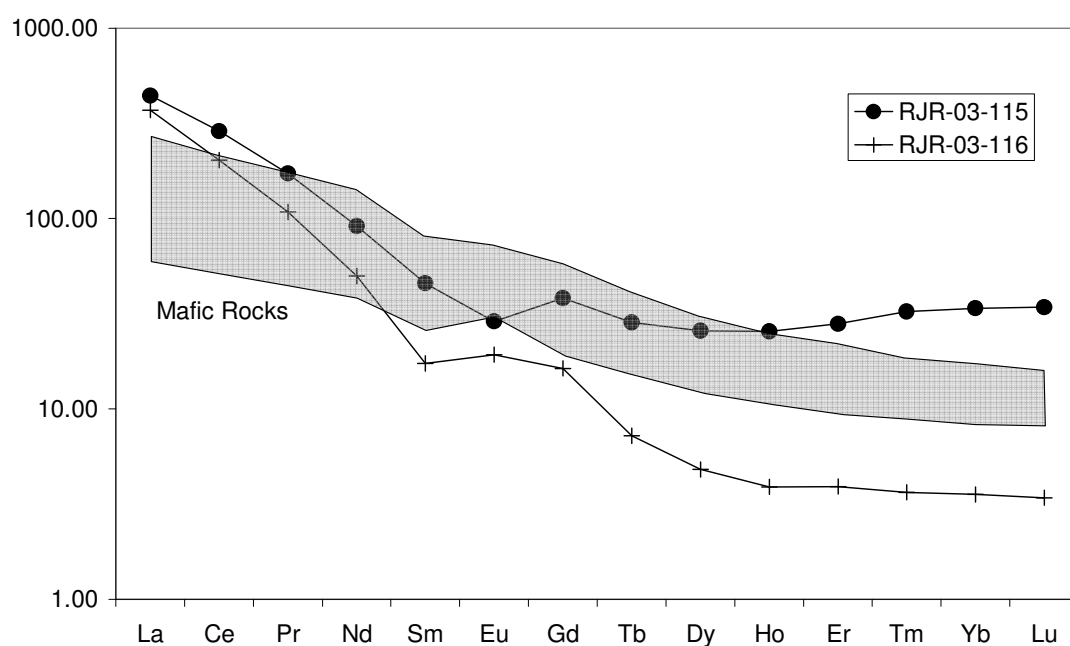


Figure 5.7- REE profiles for nepheline syenites from the SIP. Data from Table 5.2, normalised to the chondritic values of Sun & McDonough (1989)

Figure 5.7 shows the two nepheline syenites from Breivikbotn in comparison to the mafic rocks. It can clearly be seen that the profiles for both rocks differ from that of the mafic rocks, and from each other. RJR-03-115 has much higher levels of heavy REEs than RJR-03-116, an issue that needs to be discussed.

5.3.2. *Hf isotope chemistry*

As detailed in Chapter 3, the measurement of the Hf isotopic systematics in rare earth separates obtained from zircon by anion exchange chemistry has a number of advantages. Of prime importance is the abundance of Hf (0.5- 5% wt. of zircon) relative to Lu (averaging 2000 ppm in zircon), which negates the problem of having to back-calculate $^{176}\text{Hf}/^{177}\text{Hf}$ ratios to the time of formation, since at 570 Ma, the correction is less than the error on the analyses. Furthermore, a relatively simple preparation is all that is required to remove potential interferences on the Hf signal obtained by mass spectrometry, ensuring accurate measurement of the signal (e.g. Patchett & Tatsumoto, 1980).

Owing to the extensive U-Pb geochronological programme undertaken for this study (Chapter 4), a large number of samples were available for $^{176}\text{Hf}/^{177}\text{Hf}$ analysis. Table 5.4 lists the samples and their measured $^{176}\text{Hf}/^{177}\text{Hf}$ ratios, together with an ϵ_{Hf} value calculated at the age of the rocks. Geochemical data for some of these rocks are not available, but these rocks are included in this table for completeness.

Table 5.4. $^{176}\text{Hf}/^{177}\text{Hf}$ ratios measured in samples from the SIP

Zircon No.	Sample	Rock Type	$^{176}\text{Hf}/^{177}\text{Hf}$	2 σ error	Age (Ma)	2 σ error	ϵ_{Hf}	Error
Mafic Rocks								
63/54	RJR02-3d	Gabbro	0.282551	0.000006	565	9	3.5	0.2
52/14	RJR02-3d	Gabbro	0.282560	0.000007	565	9	3.8	0.3
74/57	RJR02-29i	Gabbro	0.282429	0.000002	562	6	-0.9	0.1
74/60	RJR02-29i	Gabbro	0.282424	0.000003	562	6	-1.1	0.1
74/55	RJR03-129c	Gabbroonorite	0.282564	0.000008	569	9	4.1	0.3
83/62	RJR03-129d	Gabbroonorite	0.282351	0.000004	566	1	-3.6	0.1
83/60	RJR03-129d	Gabbroonorite	0.282391	0.000080	566	1	-2.1	2.8
Intermediate								
51/1	RJR03-40b	Monzodiorite	0.282522	0.000003	566	4	2.5	0.2
51/2	RJR03-40b	Monzodiorite	0.282509	0.000004	566	4	2.0	0.3
63/59	RJR02-40b	Monzodiorite	0.282519	0.000004	566	4	2.4	0.3
63/60	RJR02-41c	Monzonite	0.282598	0.000004	565	5	5.2	0.3
63/51	RJR02-35	Diorite	0.282586	0.000004	571	4	4.9	0.3
66/54	RJR02-8	Monzonite	0.282477	0.000006	560	5	0.8	0.4
66/55	RJR02-8	Monzonite	0.282452	0.000004	560	5	-0.1	0.3
66/56	RJR02-8	Monzonite	0.282456	0.000012	560	5	0.0	0.9
Alkaline Rocks								
98/6	RJR04-245	Nepheline Syenite	0.282507	0.000007	579	14	2.3	0.2
98/7	RJR04-246	Nepheline Syenite	0.282626	0.000003	570	2	6.3	0.1
98/8	RJR04-246	Nepheline Syenite	0.282626	0.000005	570	2	6.3	0.2
98/9	RJR04-246	Nepheline Syenite	0.282637	0.000009	570	2	6.7	0.3
95/26	RJR03-116	Nepheline Syenite	0.282648	0.000005	570	2	7.0	0.2
66/57	RJR03-116	Nepheline Syenite	0.282648	0.000005	570	2	7.0	0.2
66/60	RJR03-116	Nepheline Syenite	0.282650	0.000006	570	2	7.1	0.2
66/62	RJR03-116	Nepheline Syenite	0.282638	0.000005	570	2	6.7	0.2
75/60	RJR03-116	Nepheline Syenite	0.282668	0.000003	570	2	7.8	0.1
75/61	RJR03-116	Nepheline Syenite	0.282634	0.000002	570	2	6.6	0.1
75/58	RJR03-116	Nepheline Syenite	0.282635	0.000003	570	2	6.6	0.1
81/20	RJR02-34e	Carbonatite	0.282588	0.000005	574	5	5.0	0.2
81/21	RJR02-34e	Carbonatite	0.282582	0.000005	574	5	4.8	0.2
Granitic Rocks								
74/54	RJR03-129b	Granite	0.282323	0.000003	561	4	-4.7	0.2
78/54	RJR03-129b	Granite	0.282469	0.000008	561	4	0.5	0.5
52/13	RJR02-37a	Granite	0.282374	0.000013	563	2	-2.8	0.9

Note: Measured $^{176}\text{Hf}/^{177}\text{Hf}$ ratios are not corrected for ^{176}Lu . ϵ_{Hf} values are calculated at the age of the rock. Values for present day CHUR are $^{176}\text{Hf}/^{177}\text{Hf} = 0.282818$, $^{176}\text{Lu}/^{177}\text{Hf} = 0.332$, with a decay constant for ^{176}Lu of 1.94×10^{-11} (Workman & Hart, 2005).

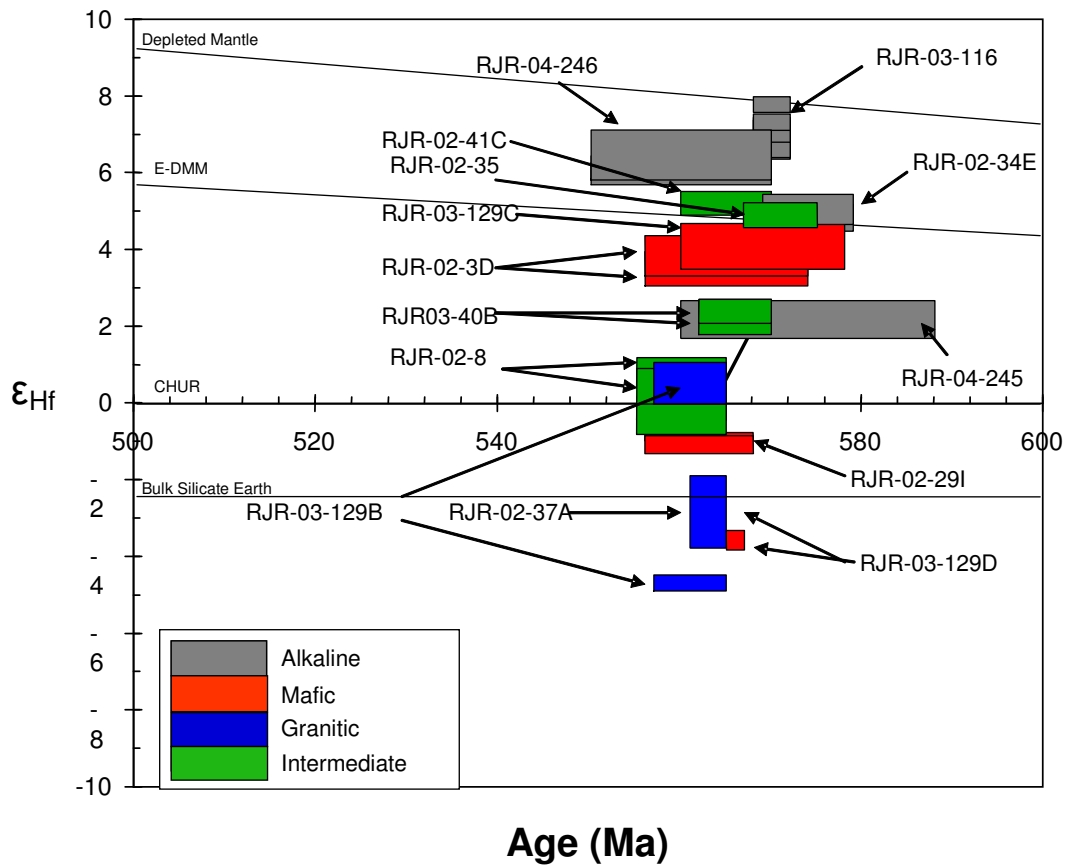


Figure 5.8- The ϵ_{Hf} variation amongst the rocks of the SIP. Data from Table 5.4, with curves for the Depleted Mantle, Enriched Depleted Morb-like Mantle (E-DMM) obtained from values in Workman & Hart (2005).

Figure 5.8 displays the variation in the measured $^{176}\text{Hf}/^{177}\text{Hf}$ rocks amongst the rocks of the SIP. The most primitive ratios are those obtained from nepheline syenite RJR-03-116 and another syenite from Breivikbotn, RJR-03-246. The rest of the rocks form a continuum from $\epsilon_{\text{Hf}} = +8$ to $\epsilon_{\text{Hf}} = -5$. It can be seen that granitoid RJR-02-37A and granite RJR-03-129B have the lowest ϵ_{Hf} values. It should be noted that one portion of zircon from RJR-03-129B plots very differently to the other portion sampled. This can be explained with reference to the U-Pb systematics of this granitoid, which revealed that a large proportion of the zircons (and thus the Hf) recovered from the rock was inherited from a much older protolith. Thus, the variation in this sample reflects a

varying amount of inherited material. Furthermore, the monzonitic rocks collected from the Øksfjord peninsula (RJR-02-40B and 41C) and the diorite from Breivikbotn (RJR-02-35) have ϵ_{Hf} values similar to or higher than those of the mafic rocks with which they are interlayered.

5.3.3. *Sm-Nd geochemistry*

Sm-Nd geochronological work has been conducted on many rocks in the SIP, and includes rocks from areas not included in this study (Daly et al., 1991; Mørk & Stabel, 1991; Cadow 1993). A comprehensive Sm-Nd and Rb-Sr study was also conducted on the Hasvik pluton by Tegner et al. (1999). It is possible to recalculate this previous work to the new ages provided in Chapter 4, to detail the $^{143}\text{Nd}/^{144}\text{Nd}$ character of the SIP rocks. Furthermore, a small range of rocks were analysed for Sm-Nd isotope chemistry during this study, including granitoids and nepheline syenites not analysed by previous studies. These analyses, together with previous work and ages for the different rocks, are recorded on Table 5.5.

These data show a wide variation in calculated ϵ_{Nd} values (Figure 5.9). The most primitive rocks are the nepheline syenites from Breivikbotn (RJR-03-115 and 116), and gabbros collected from the island of Stjernøy (Daly et al., 1991; Cadow 1993), with ϵ_{Nd} values range from +4 to +5.1 for these rocks.

Table 5.5. Sm-Nd data for rocks in the SIP

Rock	Type	¹⁴³ Nd/ ¹⁴⁴ Nd Today	¹⁴⁷ Sm/ ¹⁴⁴ Nd Today	Age (Ma)	¹⁴³ Nd/ ¹⁴⁴ Nd Initial	ε _{Nd}
<i>This Study</i>						
RJR-03-115	Nepheline syenite	0.512450	0.0856	570 ± 2	0.512130	4.4
RJR-03-116	Nepheline syenite	0.512329	0.0588	570 ± 2	0.512109	4.0
RJR-02-3B	Gabbro	0.512366	0.1002	565 ± 9	0.511995	1.7
RJR-03-125	Gabbro	0.512423	0.1316	565 ± 9	0.511936	0.5
RJR-03-129A	Granite	0.511986	0.1179	561 ± 4	0.511553	-7.1
RJR-03-129B	Granite	0.512022	0.1139	561 ± 4	0.511603	-6.1
RJR-03-129C	Gabbro	0.512318	0.1295	569 ± 9	0.511835	-1.4
RJR-03-129D	Orthopyroxenite	0.512206	0.1253	566 ± 1	0.511742	-3.3
<i>Mørk & Stabel (1991)</i>						
o31	Gabbro	0.512781	0.1904	560 ± 10	0.512095	3.2
o32	Gabbro	0.512746	0.1740	560 ± 10	0.512119	3.7
o43	Gabbro	0.512765	0.1826	560 ± 10	0.512107	3.5
033a	Gabbro	0.512723	0.1606	560 ± 10	0.512144	4.2
033b	Gabbro	0.512686	0.1608	560 ± 10	0.512107	3.5
o42	Gabbro	0.512768	0.1685	560 ± 10	0.512161	4.5
028a	Gabbro	0.512432	0.1153	569 ± 9	0.512002	1.9
o9	Gabbro	0.512422	0.1338	569 ± 9	0.511923	0.4
<i>Cadow (1993)</i>						
LAC20RC	Gabbro	0.512508	0.0985	569 ± 9	0.512141	4.6
LACF40KK	Gabbro	0.512655	0.1417	569 ± 9	0.512127	4.3
LAC102RC	Gabbro	0.512693	0.1417	569 ± 9	0.512165	5.1
<i>Daly et al. (1991)</i>						
oks3wr	Gabbro	0.512529	0.1301	569 ± 9	0.512044	2.7
sja.kv1wr	Gabbro	0.512812	0.1697	569 ± 9	0.512179	5.4
<i>Tegner et al. (1999)</i>						
Hasvik lower zone	Gabbro	0.512689	0.1604	562 ± 6	0.511973	3.6
Hasvik Upper Zone	Gabbro	0.512367	0.1739	562 ± 6	0.511727	-3.7
Contaminant	Quartzite	0.512528	0.1671	562 ± 6	0.511528	-7.5

Note: Values for present day CHUR is taken as ¹⁴⁴Nd/¹⁴³Nd= 0.512638, ¹⁴⁷Sm/¹⁴⁴Nd= 0.1967, and a decay constant of 6.54×10^{-12} (DePaolo & Wasserburg, 1976)

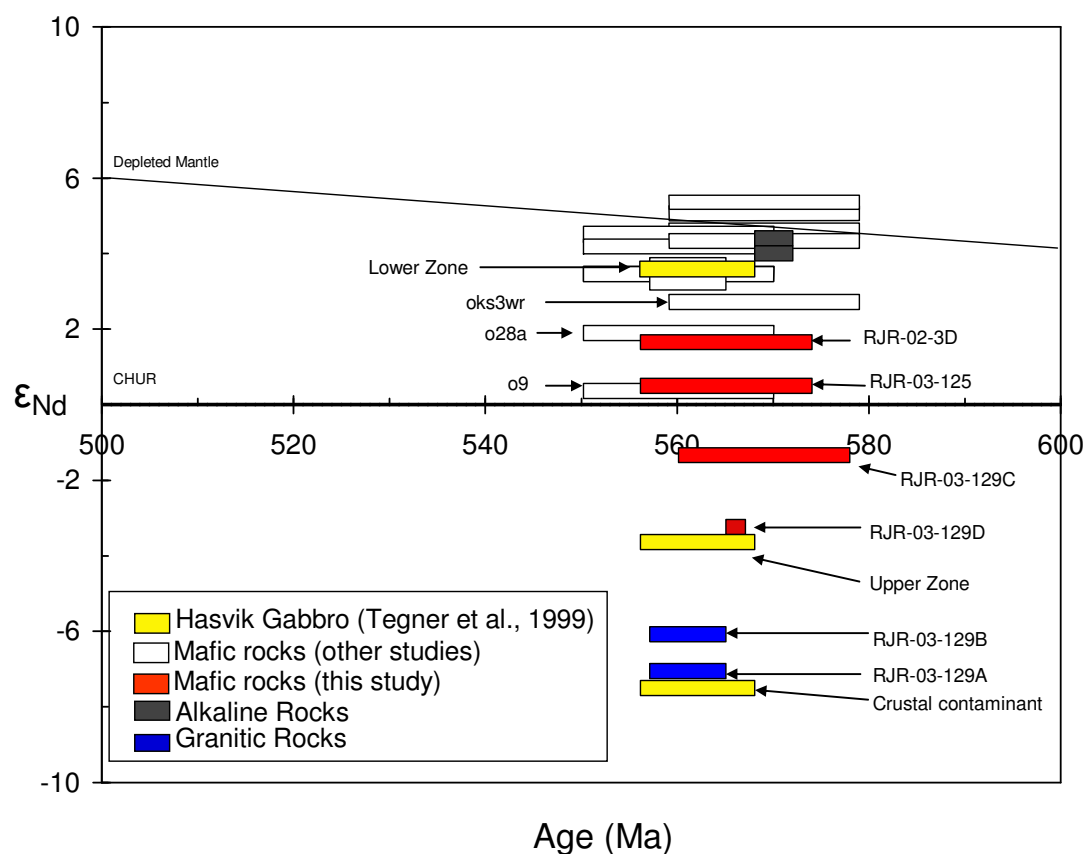


Figure 5.9- ϵ_{Nd} values for the rocks of the SIP, recalculated to the age of the rock. Depleted mantle curve from Workman & Hart (2005), data from table 5.5.

It can also be observed that many of the gabbros in the SIP display a large range of ϵ_{Nd} values, ranging from +3 to -4. The data from the Hasvik Gabbro, as provided by Tegner et al. (1999), show a range from +4 to -4, which has been attributed to variable crustal contamination. The most evolved rocks in the sequence are the granites RJR-03-129B and RJR-03-129A, which display what would normally be considered crustal signatures of $\epsilon_{Nd} = -6$ and -7 , respectively.

5.4. Interpretation of results

5.4.1. Major element geochemistry of the SIP mafic rocks

The major element chemistry of a suite of mafic rocks is generally representative of the mineralogical composition of the rock samples, and thus changes with the mineral composition of the rock. As such, the bulk chemistry of the starting magma is of great importance, and different magmas yield different crystallisation sequences depending on the initial magma composition and processes such as magma mixing and contamination. However, this is only true for rocks which have crystallised in-situ, since rocks comprised of cumulus minerals are effectively collections of minerals formed at disparate times and positions in the crystallising liquid. In the absence of known parental liquid compositions for the rocks of the SIP, it is useful to investigate the major element variations present to assess the possible variability of the starting magmas for different plutons.

Various elements can be considered in relation to the question of multiple starting magma compositions for the different plutons of the SIP. Figure 5.10 shows the CaO percentage in the rocks plotted against the Mg# ($\text{Mg}/(\text{Mg}+\text{Fe})$) for the different rocks. In this diagram, the stratigraphic trend for the Hasvik Gabbro is plotted, showing how the cumulus rocks in the layered series in that pluton evolve from the Lower Zone, through the Main Zone to the Upper Zone, possibly paralleling the fractionation trend in the original liquid (Tegner et al., 1999). Also shown is the Rognsund Intrusion, divided into the Contaminated Zone and the Layered Sequence. In general, the three data sets plot very differently in this diagram- the Hasvik Gabbro has generally higher Mg numbers than the Rognsund gabbro or the rocks analysed for this study. Furthermore, the rocks analysed in this study are lower in CaO than the Rognsund

rocks, but do show some overlap with both the Rognsund Contaminated and Layered Series rocks.

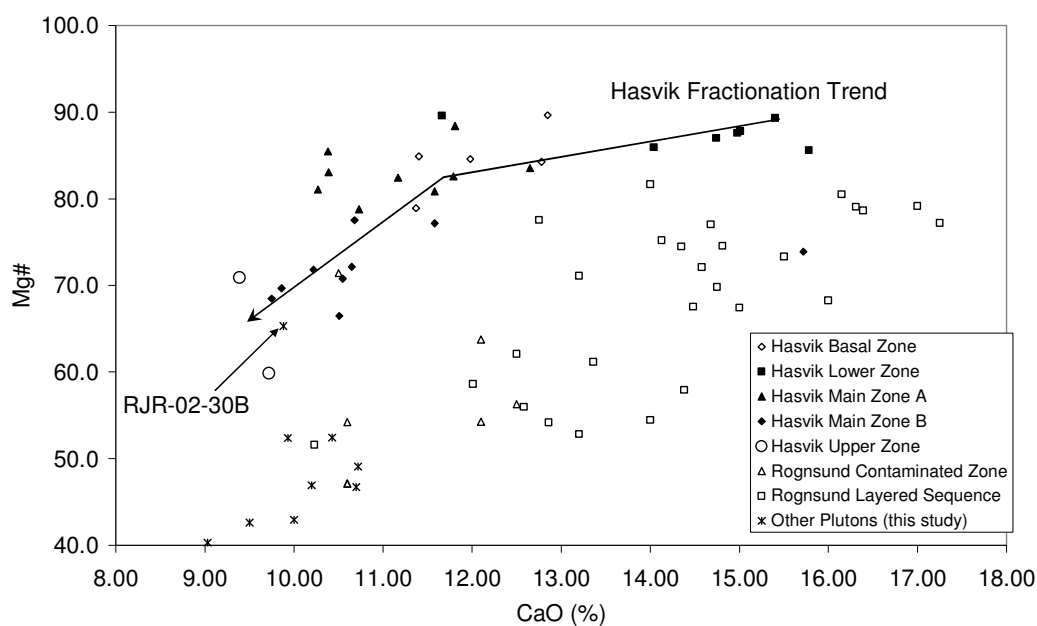


Figure 5.10- CaO (%) versus Mg#, for gabbroic samples from Table 5.1. Additional samples from Hasvik Gabbro (Tegner et al., 1999) and the Rognsund Gabbro (Robins, 1982).

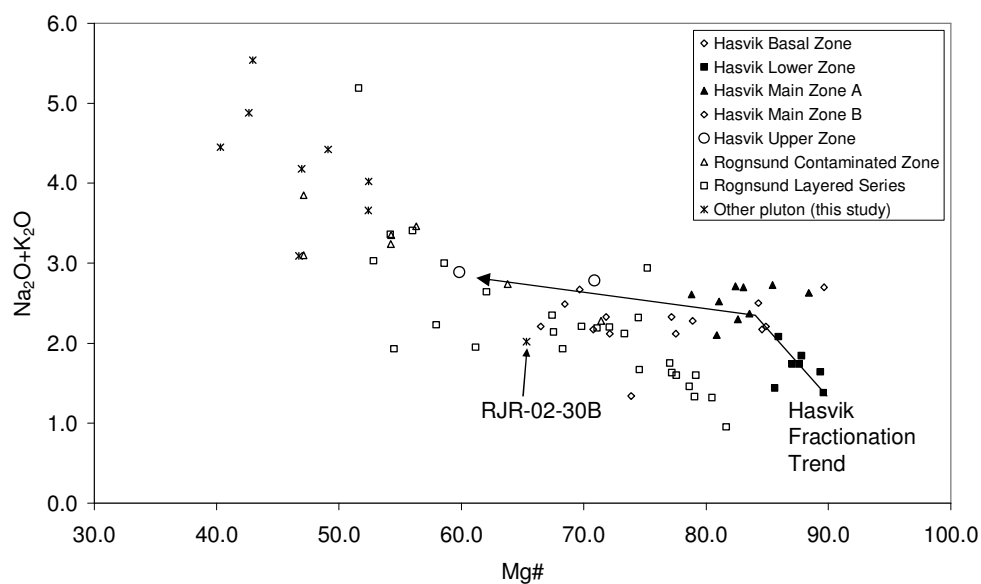


Figure 5.11- Mg# versus Na₂O + K₂O, for gabbroic samples from Table 5.1. Additional samples from Hasvik Gabbro (Tegner et al., 1999) and the Rognsund Gabbro (Robins, 1982).

Mg# versus alkali content for the gabbroic rocks of the SIP is considered in Figure 5.11. It can clearly be seen that the rocks sampled in this study are high in alkalis and low in Mg compared to the detailed studies of the two gabbros, although the one rock in overlap between the studies (sample RJR-02-30B from the Upper Zone of the Hasvik Gabbro) plots amongst the other Upper Zone rocks from this study.

It is clear from these observations that the sample set obtained in this study is not as representative as could be hoped. Using the classification scheme of Robins & Gardner (1975), the sample set has only one example of a tholeiitic gabbro (RJR-02-30B) and no examples of clinopyroxene gabbros. Although Robins & Gardner (1974) offer no firm geochemical definitions for their rock classification scheme, it is likely that the gabbros studied in this scheme would all be classified as “syenogabbros”. A better term for these rocks would be evolved high-alkali gabbros, since they contain relatively high values of alkali elements and are enriched in Fe compared to Mg. This term is used in the rest of the discussion.

5.4.2. The trace element variation amongst the mafic rocks of the SIP

It has already been noted that amongst the rocks sampled in this study, there is only limited trace element variation. This limited variability can also clearly be seen in the REE profiles for the studied rocks. Another approach to investigating possible linear relationships between pairs of trace elements is to consider the behaviour of ratios of elements. Figure 5.12 shows a variety of these ratio plots, wherein the ratio of two incompatible elements is graphed against the concentration of the less compatible of the two elements. It can be seen in these plots that none of the data shows a clear trend, although the rocks generally behave similarly. Importantly, it can be observed

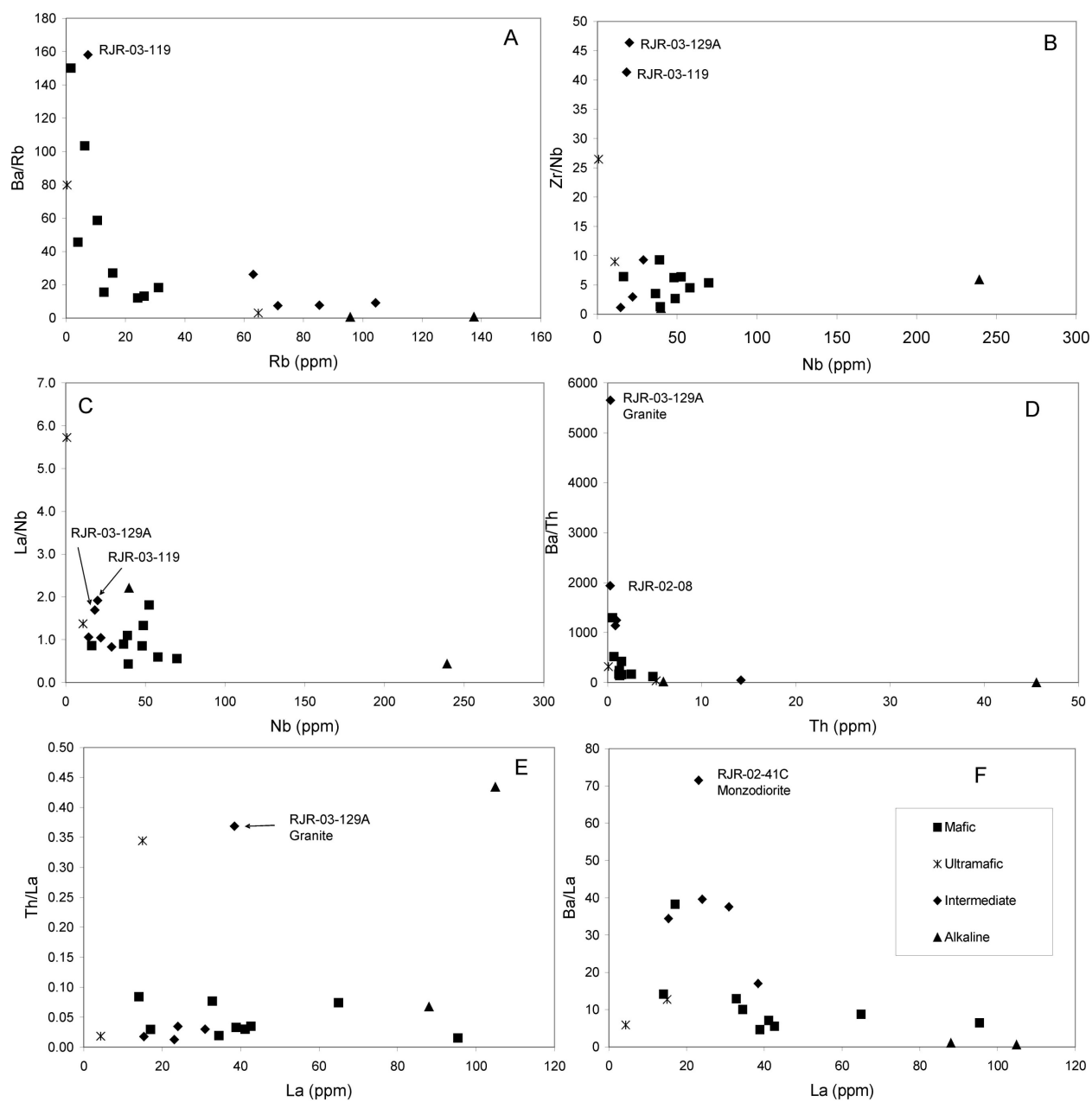


Figure 5.12- Plots of elemental ratios versus incompatible elements for rocks from the SIP. Data from Table 5.2.

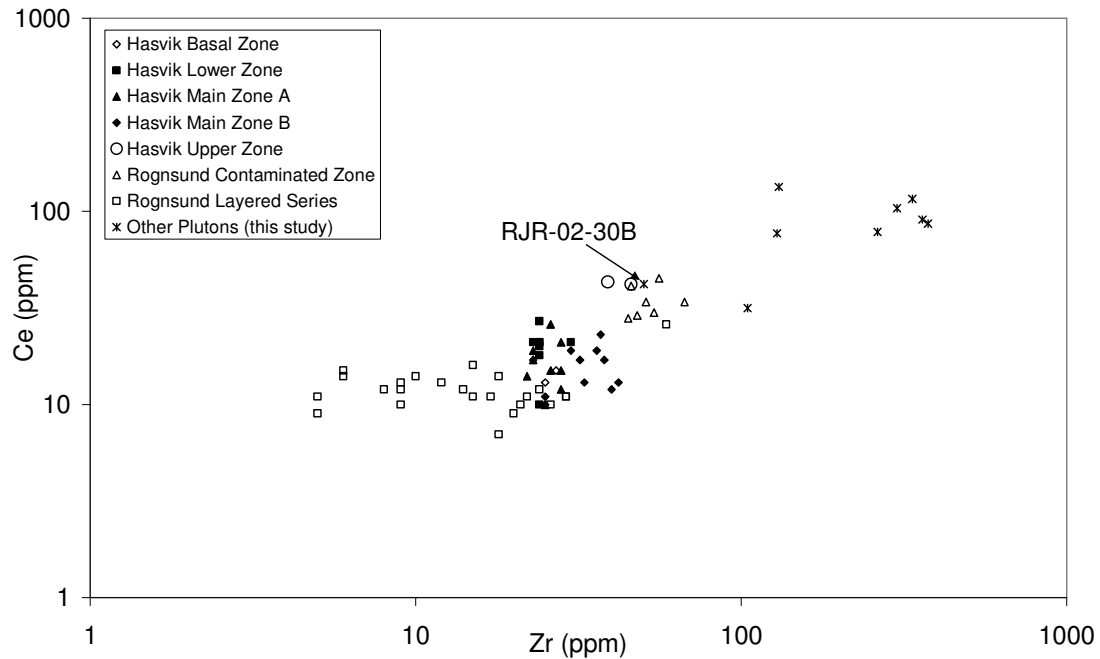


Figure 5.13- Zr versus Ce for gabbroic samples from Table 5.1. Additional samples from Hasvik Gabbro (Tegner et al., 1999) and the Rognsund Gabbro (Robins, 1982).

that the Ba/Rb ratio for the SIP gabbros is extremely variable. A similar observation can be made for Th, and one gabbro, RJR-03-125, shows an abnormal Ba/La ratio. Rb and Ba are elements that can be mobile during metamorphism, and it is possible that these elements have been lost from some of the SIP rocks, an observation supported by the re-setting of the Rb-Sr isotopic system in many plutons of the SIP (Roberts et al., 2006).

Ideally, this data set should be compared with trace element data from the tholeiitic and clinopyroxene gabbros from the SIP. However, only a very limited set of trace elements is available for the other gabbros. Figure 5.13 shows the variation of Zr and Ce within the rocks of the SIP, incorporating data from the Hasvik Gabbro (Tegner et al. (1999) and the Rognsund Gabbro (Robins, 1980). It can be seen in Figure 5.13 that the high alkali gabbros from this study are enriched in both Zr and Ce compared to the two gabbros from the literature. However, the limited data sets from the other

gabbroic suites make it difficult to extend the conclusions from this figure to the SIP as a whole.

5.4.3. Comparison with REE geochemistry of other layered intrusions

Ideally, the REE geochemistry recorded in cumulus rocks can be linked to the enrichment of trace elements in the liquid trapped in the intercumulus spaces between minerals. However, this approach requires knowledge of the original igneous petrography present in the samples to be investigated. In the case of the SIP gabbros in this study, that information is not available, having been obscured by the growth of decompression-related symplectites and recrystallisation of plagioclase (Chapter 2, Appendix A). All of the rocks in this study have experienced amphibolite facies metamorphism, as can clearly be seen in the photomicrographs in Appendix A. As such, it is very difficult to ascertain the original ratio of cumulus minerals to intercumulus space in these rocks. Since most highly incompatible elements, including the rare earths, are concentrated into the intercumulus fluid prior to crystallisation, an estimate of such is required for modelling the trace element evolution. This lack of a control on the percentage of intercumulus liquid in the original cumulus rocks means that any form of trace element modelling is necessarily unconstrained and inconsistent.

In the absence of other gabbros from the SIP to which one can compare the trace element data, and the lack of suitable petrography, it is instructive to compare the trace element chemistry of the SIP rocks with other igneous intrusions. It has been stated above that the high alkali SIP gabbros have extremely high concentrations of REEs compared to other gabbroic intrusions enriched, and this point is emphasised

when comparing the SIP gabbros to other layered intrusions. A selection of rocks from three different layered intrusions has been plotted in Figure 5.14 for comparison with the SIP gabbros. In Figure 5.14A, eight rocks from the Main and Upper Critical Zones of the Bushveld Complex in South Africa are plotted (Maier & Barnes, 1998). The Bushveld Complex is the largest layered intrusion known on Earth, and was formed through the repeated injections of fresh magma into the magma chamber (Eales & Cawthorn, 1996). Though the exact degree of crustal contamination involved in the formation of the Bushveld Complex is unknown, it is generally thought to be quite high (Harris et al., 2005). In Figure 5.14B, the REE profiles from parts of the extensively studied Skaergaard Complex in East Greenland are presented (McBirney, 2002). Skaergaard, although relatively small, has been shown to have experienced measurable crustal contamination (Stewart & De Paolo, 1990), and also features the Sandwich Horizon, a zone wherein the remaining, incompatible-enriched melt left by bottom-up and top-down crystallisation was trapped. Figure 5.14C shows the REE profiles from the Bjerkreim-Sokndal Intrusion in southern Norway (Charlier et al., 2005). Not only is the parental jotunite magma for this intrusion theorised to be enriched in trace elements, the intrusion itself has undergone crustal contamination (Tegner et al., 2005).

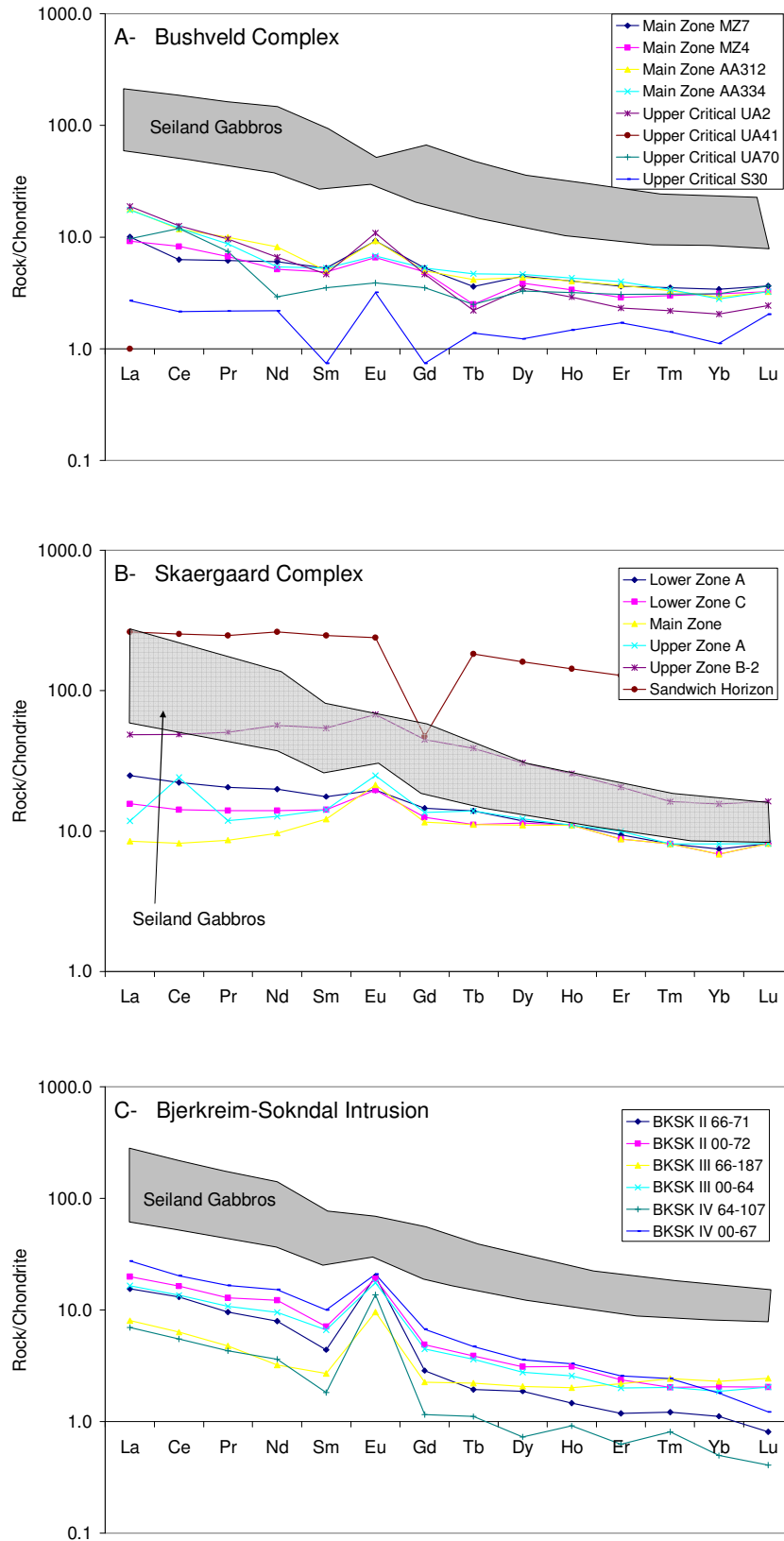


Figure 5.15- REE Profiles from selected layered intrusions. A- Bushveld Complex, data from Maier & Barnes (1998) with some data extrapolated, B- Skaergaard Complex, data from McBirney (2002), C- Bjerkreim-Sokndal Intrusion, data from Charlier et al. (2005).

It is immediately clear on examination of these profiles that only two rocks, from the Sandwich Horizon and the top of Upper Zone in the Skaergaard Intrusion, have REE abundances that reach or exceed those of the SIP gabbros. The rest of the rocks have REE abundances lower than those of the SIP rocks. Furthermore, the Skaergaard and Bushveld Complexes show considerably more relative variation in their REE contents than the SIP gabbros. If one considers the lowest La value reported for each intrusion compared to the highest La value for each intrusion, the Bushveld shows a 10-fold variation, the Skaergaard Complex a 30-fold variation, the Bjerkreim-Sokndal Intrusion a 3-fold variation and the SIP gabbros a 4-fold variation.

If the SIP gabbros are enriched in REEs relative to other layered intrusions, then it is perhaps appropriate to compare their trace element abundances not only with layered intrusions, but with ocean island basalts (OIB) which are enriched compared to the mid-ocean ridge basalts (MORB). The liquids parental to many continental layered intrusions are often considered to be similar to those parental to MORB (e.g. Halama et al., 2004; Tegner et al., 2005). Halliday et al. (1995) and Allègre et al. (1995) presented large data of OIB analyses, and Allègre et al. (1995) also presented a set of averaged MORB analyses. Although the REE dataset for these samples is minimal, it is of interest to plot the SIP gabbros alongside these oceanic basalts for a variety of elements, as can be seen in Figure 5.15.

In this figure, it can be noted that the SIP gabbros generally plot in the same field as the OIBs, rather than with MORB or the other layered intrusions. However, it can be shown that the Rb and Th data from some SIP gabbros are often lower than the

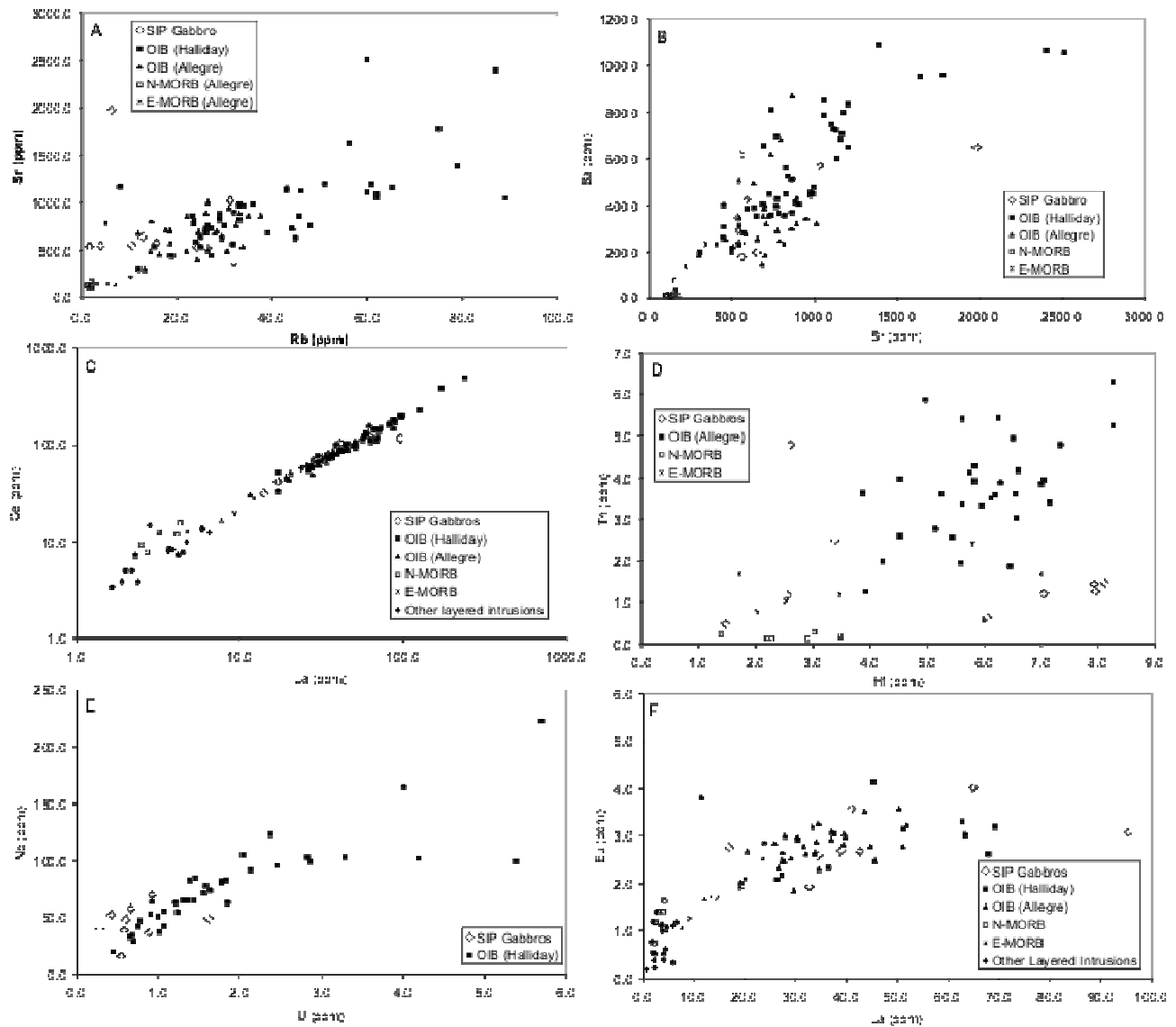


Figure 5.15- Comparison of SIP gabbros with oceanic basalts and layered intrusions. OIB data are taken from Halliday et al. (1995) and Allègre et al. (1995), MORB data are taken from Allègre et al., (1995), and layered intrusion data are taken from Charlier et al (2005), Maier & Barnes (1998) and McBirney (2002).

general OIB trend, which may indicate that these elements have been removed from the SIP rocks during metamorphism. Tying all these threads together, it is possible to confirm that the SIP gabbros analysed in this study have trace element abundances which are anomalous compared to other gabbroic sequences in well-known layered intrusions such as Skaergaard and the Bushveld. Owing to the over-representation of the high alkali gabbros in the current data set, this conclusion can only tentatively be

extended to the whole SIP. However, it should be remembered that the SIP gabbros are spatially and temporally associated with relatively large volumes of alkaline magmatism, including nepheline syenites and carbonatites, which are generally considered to be trace element enriched and sourced in volatile-enriched mantle regions, which lends credence to the analysed abundances.

5.4.4. The isotopic variation amongst the rocks of the SIP

It is clear from figures 5.8 and 5.9 that the rocks of the SIP show considerable variation in their isotopic signatures. Both the Sm-Nd and Lu-Hf systematics show that the rocks range from mantle to crustal values. In the case of the Hasvik Gabbro, for which a complete profile is available (Tegner et al., 1999), it can be seen that this variation can be modelled in terms of crustal contamination of the starting magma composition, causing a change in ϵ_{Nd} from +4 to -4. It is a logical conclusion that similar processes may have affected the other plutons in the SIP, and much of the variation in the SIP can be explained through such a mechanism.

The highest ϵ_{Hf} values sampled are present in the nepheline syenites and carbonatites of the SIP. The volatile-rich nature of these magmas makes it hard for them to assimilate any significant amount of material, and these rocks are generally considered to preserve the mantle signature of their source (Bell & Blenkinsop, 1987).

Correlating this observation with the Sm-Nd data, it can be seen that gabbros sampled from Stjernøy (Daly et al., 1991; Cadow, 1993) have ϵ_{Nd} values which overlap with the ϵ_{Nd} values for the SIP nepheline syenites. These gabbros are emplaced into pre-existing gabbro and do not abut country rocks at any point, which limits the possible crustal contamination possible for such rocks after emplacement. This is a strong

indication that the source region for these gabbros is the same as that for the nepheline syenites, and it is likely, considering the close spatial and temporal relationships between the different rocks of the SIP, that the SIP magmatism was all sourced in the same mantle region. Such a mantle region would appear to have a depleted mantle signature.

It is possible to consider the question of contamination amongst the zircons of the SIP by considering the variation in the isotopic signatures for the two isotopic signatures more carefully. Figure 5.16 shows the variation in $^{177}\text{Hf}/^{176}\text{Hf}$ relative to the variation in $1/\text{Hf}$, and figure 5.17 shows the same variation in $^{144}\text{Nd}/^{143}\text{Nd}$ versus $1/\text{Nd}$. A third figure, 5.18, shows the covariation between the two systems for the limited range of rocks for which both sets of data are available.

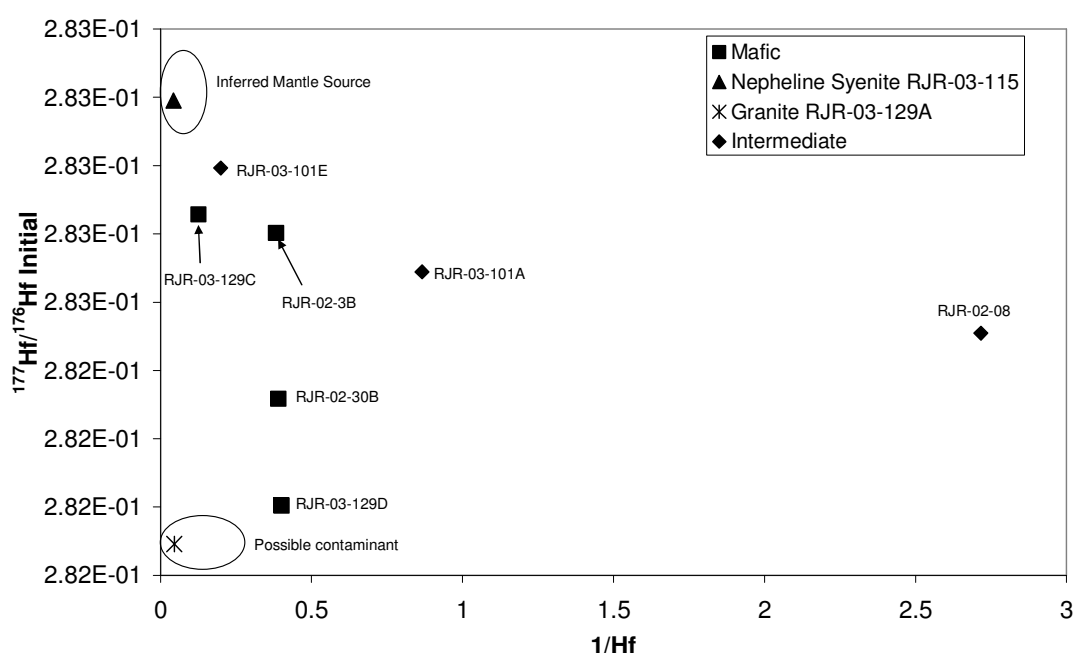


Figure 5.16- $^{177}\text{Hf}/^{176}\text{Hf}$ versus $1/\text{Hf}$ for rocks of the SIP, data from table 5.4.

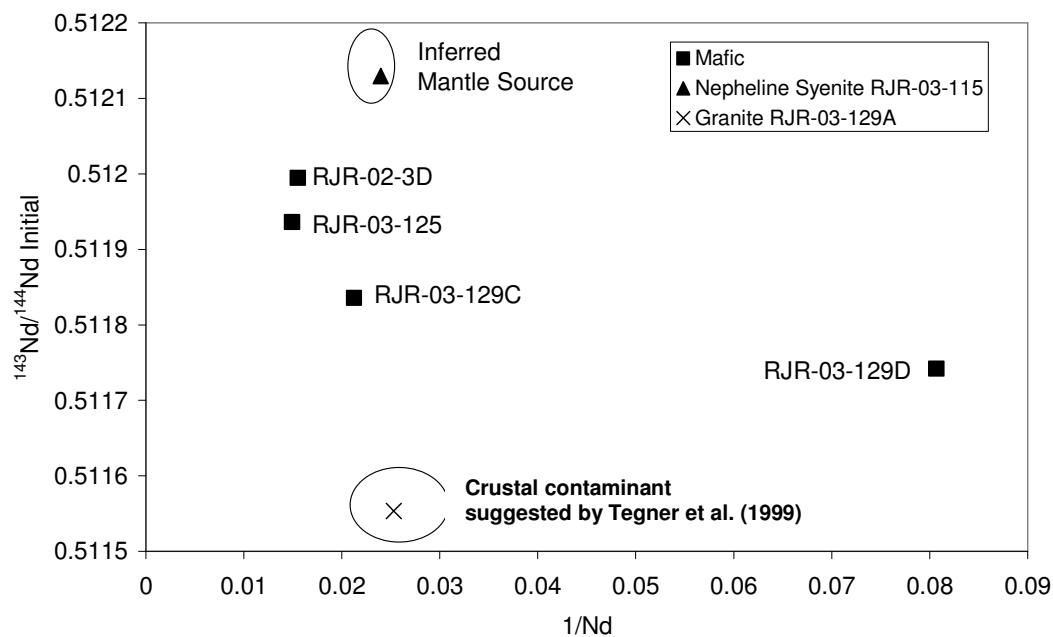


Figure 5.17- $^{144}\text{Nd}/^{143}\text{Nd}$ versus $1/\text{Nd}$ for the rocks of the SIP, data from table 5.5.

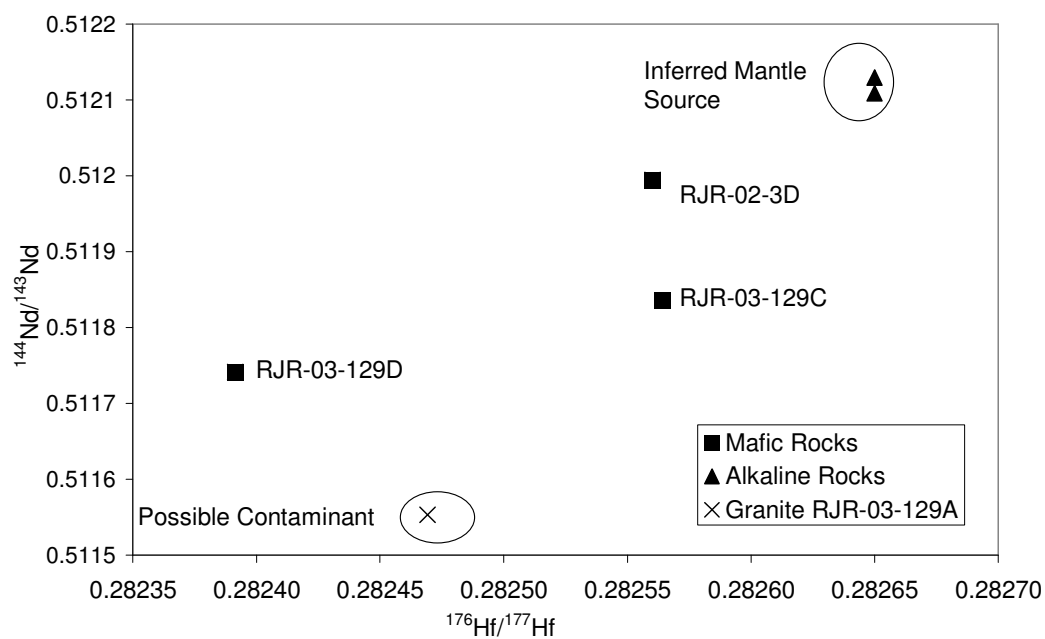


Figure 5.18- $^{177}\text{Hf}/^{176}\text{Hf}$ versus $^{144}\text{Nd}/^{143}\text{Nd}$ for rocks for rocks of the SIP.

It can clearly be seen in these diagrams that there is significant variation in both isotopic systems. In the absence of analyses from potential contaminants such as the country rocks surrounding the different intrusions, forward modelling of the AFC

processes in these rocks was found to be unconstrained, owing to the lack of information on potential contaminants. However, the forward modelling clearly indicated that a single set of AFC parameters, such as that present by Tegner et al. (1999), is insufficient to explain the variation present in the SIP, and it is also clear from the figures above that a single contaminant as suggested by Tegner et al. (1999) is unable to account for the variation in rocks such as RJR-03-129D. Thus, in the treatment of the isotopic data below, a qualitative approach to the data was taken.

Tegner et al. (1999) suggested one possible contaminant for the Hasvik Gabbro. This contaminant has a $^{144}\text{Nd}/^{143}\text{Nd}$ value close to that of granite RJR-03-129A, and this granitoid is used to extrapolate the suggested contaminant to all three diagrams. The nepheline syenites are taken to represent the isotopic composition of a possible mantle source. If one considers the assimilation-fractional contamination (AFC) projections given by De Paolo (1981) relative to these diagrams, it can be seen that the suggested contaminant cannot be the only contaminant involved in the contamination of the SIP rocks. It is especially apparent that RJR-03-129D, an orthopyroxenite, which plots close to the suggested contaminant, would have to comprise primarily crustal material in order to satisfy the demands of AFC calculations using the contaminant suggested by Tegner et al. (1999). If AFC processes are to be the driving mechanism behind the variation amongst the SIP gabbros, then several contaminants are likely to be involved.

Another important conclusion to draw from the isotopic data is that not all the gabbros investigated in this study have isotopic signatures indicative of considerable crustal

contamination. This has implications for the interpretation of the trace element profiles detailed above.

5.4.5. The relationship between the intermediate and the mafic rocks of the SIP

Intermediate, feldspar- and quartz- rich igneous rocks such as monzonites and diorites are generally considered to be the products of the melting of crustal materials, rather than the direct products of mantle-derived magmas. The separation of significant quantities of felsic melt from a mafic melt is unfeasible on petrologic grounds, and a mafic melt can be expected to produce less than 5% felsic melt by volume by fractionation (Hess, 1985). Thus, it should be expected that in the case of the SIP, where 10% of the rocks are felsic, the felsic rocks associated with the SIP would be crystallised from crustally derived melts.

This is patently not the case in the dataset presented. Not only are most of the monzonites and diorites sampled indistinguishable from the mafic rocks in terms of REE profiles, but the intermediate rocks also have mantle isotopic signatures, commonly more primitive than the associated mafic rocks. Since these rocks are difficult to contaminate owing to their silica-rich nature, this “mantle” signature can be considered to be that of the crustal rocks from which the melt parental to the intermediate rocks was derived. These rocks can readily be contrasted to the single granite body sampled in this study, which has a crustal isotopic signature and a different REE profile to the other intermediate rocks.

Another significant feature of these rocks is the positive Eu anomaly present in most of the profiles. A rock derived from the melting of crustal material should show a

negative Eu anomaly. A positive anomaly in a mafic rock would normally be interpreted as indicative of a high cumulus plagioclase content. Although the felsic rocks are commonly rich in plagioclase, it is unlikely that significant separation and accumulation of plagioclase would occur in a high viscosity felsic melt. An alternative explanation for the positive Eu anomaly could be that the Eu anomaly represents the trace element content of the source rocks

If the intermediate rocks are not the direct products of mafic melts, then the most likely explanation for the similarities between the mafic and intermediate rocks is that the intermediate rocks are the products of melted mafic material, i.e. the intermediate rocks form through the melting of previously emplaced and solidified SIP gabbro. Such a process will preserve the mantle isotopic signal and the rare earth profile in the rocks, and it is known that the SIP gabbros were emplaced over several million years (Roberts et al., 2006). This theory does need to be confirmed by detailed petrographic and mineralogical investigation of the intermediate and the mafic rocks, and especially of the plagioclase minerals present in both rock types.

5.5. Discussion

The data presented here indicate that the high alkali gabbros in the SIP are enriched in trace elements relative to other mafic intrusions, and contain levels of trace elements more indicative of ocean island basalts than of continental layered intrusions. In comparison, of the layered intrusions commonly considered to have been derived from depleted mantle sources, only the Sandwich Horizon and the top of the Upper Zone of the Skaergaard Complex (McBirney, 2002) have similar levels of enrichment in the REEs and other trace elements. These represent, respectively, the repository for

late-stage, highly evolved magmatic fluid enriched in incompatible elements, and a heavily contaminated zone in which up to 20% by mass of the rock present is derived from sedimentary sources (Stewart and De Paolo, 1990).

In contrast, the dataset from the SIP comprises samples from several spatially distinct plutons, with most samples displaying the growth of spinel-orthopyroxene symplectites after olivine, indicative of an olivine-bearing mineral assemblage. These rocks do not contain significant apatite or zircon that would indicate that they were formed from a highly differentiated magma, and so are not similar to the Sandwich Horizon in Skaergaard. Furthermore, the homogeneity of the SIP gabbros argues against high degrees of fractional crystallisation. It has been observed that in the limited datasets consulted from both Skaergaard and the Bushveld that the variation in REE content in both these complexes is significantly more than amongst the SIP gabbros, each of which can be expected to have undergone its own, unique magmatic evolution after emplacement.

It should be noted that the trace element contents of cumulus rocks is directly related to the percentage of melt trapped in the intercumulus spaces within the rock. Thus, it is possible that the enriched nature of the SIP rocks could be related to a high proportion of trapped liquid in the cumulates of the SIP plutons. This is difficult to ascertain on the mineral assemblages currently present in the rocks, as these assemblages have experienced significant metamorphism since emplacement. However, it is unlikely that all the rocks in this study, derived from numerous different plutons, would all contain similar amounts of trapped liquid.

If the fractional crystallisation of a magma derived from the depleted mantle is insufficient to explain the enrichment of the SIP gabbros, it is appropriate to consider whether the crustal contamination of mantle-derived magmas would have resulted in the trace element enrichment present in the rocks. Crustal rocks are generally enriched in incompatible trace elements and REEs, and Tegner et al. (1999) postulated that at least one pluton in the SIP, the Hasvik Intrusion, had undergone significant crustal contamination. De Paolo (1981) pointed out that while a magma generally has sufficient energy from the latent heat of crystallisation to melt a large portion of country rock, its ability to assimilate large quantities of such melt is limited by a variety of factors such as the size of the intrusion and its surface area, which determine the rate of assimilation and the rate of cooling. As such, it can be argued that if the same magma was emplaced into a variety of differently shaped chambers in different volumes, the amount of assimilation-fractional crystallisation (AFC) each magma chamber will undergo may differ greatly (e.g. Davidson et al., 2002). Tegner et al. (2005) noted that between three different essentially basaltic gabbroic intrusions, radically different rates of assimilation and volumes of assimilated material were present.

The homogeneity of the REE element data from the SIP does not fit comfortably with the isotopic variation observed in the rocks of the SIP. Whereas the REE content of the high alkali SIP gabbros is extremely uniform, the isotopic ratios in the SIP gabbros vary greatly, from mantle values to crustal values. Tegner et al. (1999) noted such large variations within a single pluton in the SIP, and ascribed the process to progressive contamination of an essentially basaltic magma derived from the depleted mantle. A similar range of isotopic values has been noted amongst chemically similar,

but isotopically diverse ocean island basalts (e.g. Halliday et al., 1995; Allegre et al., 1995), and ascribed to differences in the isotopic signatures between otherwise homogenous mantle sources. Both of these hypotheses need to be considered in the light of the data presented here.

The idea that the plutons of the SIP are derived from a wide variety of isotopically diverse mantle sources similar to ocean island basalts is not considered reasonable, since all the plutons are constrained to a very limited area, and any magmas derived from the mantle below the SIP can reasonably be expected to have interacted and re-equilibrated with the surrounding mantle rock during their ascent (O'Hara & Herzberg, 2002). The isotopic variation amongst the SIP gabbros is much more readily explained by positing a single depleted mantle source, as evidenced by the overlapping isotopic values for the alkaline rocks from Breivikbotn and gabbros from Stjernøy, which has been modified by assimilation of crustal material during or after emplacement.

In a comprehensive study of the Hasvik Gabbro, Tegner et al. (1999) posited that the assimilation of a large amount of country rock adjacent to the Hasvik pluton was responsible for the contamination. The contaminant suggested by Tegner et al. (1999), a melt derived from quartzofeldspathic gneisses adjacent to the Hasvik Intrusion, will have contained abundant incompatible elements, the addition of which to a basaltic magma should be readily apparent in the REE profiles and trace element data presented above, especially in the quantity (20% of the Hasvik Intrusion by volume) suggested by Tegner et al. (1999). However, it is possible that whereas cumulus minerals crystallising from a contaminated melt retain the contaminated isotopic ratio,

the cumulates formed in the chamber do not trap sufficient liquid for the enriched trace element characteristics of the liquid to be retained in the trace element abundances in the crystalline rock. This would, however, contradict the idea that the trace element contents of the SIP rocks reflect a high percentage of trapped liquid in the original cumulus rocks. Therefore, the current data set implies that the trace element enrichment in the rocks of the SIP is a primary feature of the parental magmas, rather than a secondary feature introduced by AFC processes.

It has been noted that the dataset used in this study shows a number of differences with previously analysed data from the Hasvik and Rognsund Gabbros. This makes it difficult to compare the REE profile from the Hasvik pluton with the other profiles in the dataset. It has been noted that despite the similarities in the REE profiles of the high alkali gabbros, some of the gabbros display uncontaminated isotopic signatures. This contradicts any hypothesis in which random sampling has resulted in a data set comprising only heavily contaminated gabbros. If crustal contamination of the high alkali gabbros is present in the isotopic data but not in the trace element data, the implication is the volume of assimilated material is either small or that the contaminant is depleted in trace elements. However, if only a small volume of material is to be assimilated, then the contaminants involved in the process should have a distinctly different isotopic signature from the SIP source in order to produce the large range of isotopic variation present in the SIP gabbros. Although the Klubben Quartzite country rocks, into which the Hasvik Intrusion is emplaced, are of Proterozoic age (1.8 Ga; Kirkwood & Daly, 20034), most of the other intrusions on the Øksfjord peninsula are emplaced into the slightly older Eidvageid gneiss (Akselsen, 1982). This gneiss comprises both ortho- and paragneissic horizons, and

has been metamorphosed to granulite facies and provides a likely source for the contamination of the SIP magmas.

The enriched nature of the SIP gabbros must still be addressed. If crustal contamination was not the contributing factor to the production of such enrichment, then the enrichment must have been inherent to the original magma prior to emplacement. The limited variation amongst the REE content of the SIP gabbros is similar to that observed in ocean island basalts (Halliday et al., 1995; Allegre et al., 1995), in which case the limited variation is ascribed to a two-component lower mantle source below the garnet stability field, in which only one mineral species, pyroxene, undergoes small fractions of melting, and the melt is extracted rapidly without time to re-equilibrate with the surrounding mantle.

Furthermore, the SIP comprises numerous carbonatitic and alkaline intrusions that are generally coeval with the mafic magmatism (Chapter 4), so the mantle source area for the SIP must have been enriched in volatiles and especially CO₂ at some stage in its evolution. Mantle metasomatism is commonly mentioned in conjunction with the production of such volatile-rich magmas (Bell & Blenkinsop, 1987), and has also been posited as a possible model for the production of ocean island basalts (Pilet et al., 2005). Therefore, variable metasomatism of a lower mantle region could have produced both the alkaline rocks and the mafic rocks of the SIP, and could explain the enriched nature of the SIP gabbros. Such metasomatism could occur during the ascent of a mantle plume, but, considering the close association between carbonatites and rift zones, it is likely that the metasomatism occurred as a response to extension in either a rift or a back-arc setting.

This study has shown that the SIP magmatism is considerably variable in its geochemical character. Investigation of the variation in the sampled rocks relative to other rocks has highlighted the need for detailed geochemistry and petrography on a much wider variety of gabbroic rocks than has previously been undertaken, and the scope of this discussion is mainly limited to the high alkali suite of gabbros in the SIP. More data from not only this suite but the tholeiitic and clinopyroxene suites would greatly enhance the conclusions drawn from this limited dataset, and aid in understanding the evolution of the SIP better.

5.6. Summary of Geochemical and Isotopic Data

- 1) The Seiland Igneous Province comprises mafic, intermediate and alkaline rock types of the same age.
- 2) The mafic rocks sampled in this study appear distinct from tholeiitic and clinopyroxene gabbros previously studied in the SIP, and are considered to be high alkali gabbros, containing >3% alkalis in contrast to the <3% alkalis present in the Hasvik and Rognsund intrusions.
- 3) The mafic rocks sampled in this study from several different plutons show homogeneous REE profiles, but are isotopically diverse.
- 4) The ϵ_{Hf} and ϵ_{Nd} values for the gabbroic rocks range from +8 to -6 and from +4 to -4 respectively. Most intermediate rocks show depleted mantle isotopic values.
- 5) The gabbroic rocks are highly enriched not only in the rare earth elements but also in other trace elements. The trace element concentrations preserved in the

rocks are similar to those in ocean island basalts, and significantly higher than those in the Skaergaard, Bushveld and Bjerkreim-Sokndal intrusions.

- 6) These observations are not compatible with either a range of mantle sources or with the variable contamination of a single mantle-derived magma.
- 7) It is hypothesised that these rocks are formed from a melt produced from a metasomatised mantle source.
- 8) During emplacement, this parental melt was contaminated by the assimilation of small amounts of ancient crustal material, so as to preserve the original trace element concentrations.
- 9) Such a hypothesis is compatible with emplacement in an extensional tectonic setting such as intracontinental rifting or back-arc extension.

Chapter 6: Revisiting the metamorphic history of the SIP- towards a new tectono-metamorphic model

6.1. Introduction

Geochronology and metamorphic petrology are two complementary fields of study when dealing with areas with a long and complex history of deformation. Until this study, the age chronology for the SIP was both lacking and misleading (Chapter 1 and 4). Therefore, metamorphic studies on the area, although based on robust data, have come to conclusions as to the tectono-metamorphic history of the SIP that are inconsistent in the light of the data presented in this thesis.

In previous chapters, the following conclusions have been reached:

- The SIP was largely emplaced between 560 Ma and 570 Ma, with some alkaline magmatism occurring as late as 520- 530 Ma.
- The setting for the SIP magmatism was most likely extensional rifting, possibly in an intra-continental setting.
- There is evidence from the U-Pb zircon systematics for a metamorphic event affecting the nepheline syenites of the SIP at 420 Ma, resulting in the growth of new zircon and Pb-loss from existing zircon.
- There is no evidence for any other metamorphic events recorded in the zircon systematics.

These conclusions clash with previous interpretations of the evolution of the SIP. The oldest ideas for the evolution of the SIP posit that the SIP was emplaced sequentially

during a series of Cambrian compressional tectonic events (Sturt et al., 1978; Ramsay et al., 1985). This chronology was based on K-Ar and Rb-Sr dating, which produced ages that were too young for the SIP. Newer ideas for the SIP, based on Sm-Nd and Rb-Sr ages, hypothesise a 300 m.y. period of extensional magmatism from 850 Ma to 550 Ma (Reginiussen, 1996), after which several metamorphic events affected the SIP in the Late Precambrian and Early Cambrian periods (Elvevold & Andersen, 1993; Elvevold et al, 1994). These older models therefore need revision in the light of the new data.

This chapter does not aim to produce a comprehensive new overview for the metamorphic evolution of the SIP, but rather aims to re-examine and revise the existing metamorphic understanding of the SIP in the light of the new information gathered during the course of this study. One new set of U-Pb age data is presented, and used to disprove part of the previous metamorphic work.

6.2. Previous metamorphic work on the SIP

6.2.1. Pressure and temperature estimates from contact metamorphic aureoles.

Contact metamorphism accompanying the emplacement of the SIP is preserved in the aureoles of many plutons. Various authors working on the different plutons of the SIP have offered estimates of the temperatures and pressures during contact metamorphism. Most of the estimates are empirical and based on correlating the observed metamorphic assemblage with theoretical mineral stability curves, although some of the later work (Tegner et al., 1999) is based on direct microprobe analyses of the minerals present in the metamorphic aureoles of the plutons.

Sturt and Taylor (1971) used a variety of mineral stability curves to constrain the contact metamorphic aureole of the Storelv Gabbro to 7- 9 kbar, based on the presence of kyanite, sillimanite, cordierite, and staurolite in the gneisses adjacent to the gabbro. Bennett (1974), working with the Reinfjord Ultramafic Complex, gave a minimum pressure of 5.5 kbar for contact metamorphism, based on the co-existence of kyanite and sillimanite in the country rock. Speedyman (1983) also reports the coexistence of kyanite and sillimanite in the country rocks to the Husfjord Complex, and reports a pressure in excess of 5.5 kbar. In all these cases, only a pressure estimate is given, although temperatures required for the formation of the various mineral assemblages are in excess of 500° C (e.g. Sturt & Taylor, 1974; Bennett, 1974; Robins et al., 1990). These studies also assumed that the country rocks hosting the intrusions have not been metamorphosed before, or have been completely overprinted by the contact metamorphism. Considering the identification of much older magmatic (Daly et al., 1991) and metamorphic (Corfu et al., 2006) events than the SIP in the Kalak Nappes, this assumption is not necessarily warranted.

Robins et al. (1990) reported that the aureole of the Lille Kufjord gabbroic pluton showed two distinct metamorphic assemblages. A quartz-hercynite assemblage is overgrown by garnet and sillimanite. The early assemblage was held to represent the peak of metamorphism, at >770° C, with the later assemblage of garnet and cordierite representing the equilibrium assemblage attained during cooling, at temperatures around 650- 700°C and pressures between 5.4 and 8.2 kbar (Robins et al., 1990).

Elvevold et al. (1994) reported extensive microprobe work on contact metamorphic assemblages grown in contact-metamorphosed xenoliths within the Øksfjord gabbros,

as well as on spinel-quartz porphyroblasts within paragneisses adjacent to the gabbros. In their study, pressure and temperature during contact metamorphism accompanying the SIP magmatism are given as 5- 6.5 kbar and 930- 960° C respectively, using garnet-orthopyroxene geothermobarometry (Elvevold et al., 1994). Reginiussen, in unpublished work (1996), and in a subsequent paper (Tegner et al., 1999), reports that the aureole of the Hasvik Intrusion records a pressure of 6- 7.5 kbar with temperatures around 875°C, based on geothermobarometric calculations on a mineral assemblage including orthopyroxene, spinel, cordierite, corundum and garnet.

In general, pressure estimates for the contact metamorphism caused by the intrusion of the SIP range between 5.5 and 9 kbar, with most estimates between 5.5 and 7 kbar. This would indicate that the SIP gabbros were emplaced into the crust at a depth of between 20 and 30 km. However, the temperature estimates given by the various studies are only consistent in that most postulate a temperature in excess of 500°C. Temperature is a function of the distance from the intrusion to the contact metamorphosed rock, and the thermal conductivity of the rock itself, and different lithologies may thus record different temperatures. It should be noted that the highest estimates of temperature during contact metamorphism are obtained from xenoliths completely enclosed in mafic rock (Elvevold et al., 1994; Reginiussen, 1996), which is not unexpected considering that such xenoliths would have been exposed to the greatest heat from the intrusions for the greatest period of time.

6.2.2. Regional metamorphic studies on the SIP

Although it is generally accepted that the igneous rocks of the SIP have seen upper amphibolite facies metamorphism (Sturt et al., 1978; Ramsay et al., 1985; Tegner et al., 1999), very little work has been done on this post-emplacement phase of metamorphism. However, two studies have attempted to quantify the post-emplacement metamorphism of the SIP. Elvevold & Andersen (1993) reported fluid inclusion data from a variety of rocks within the SIP. Elvevold et al. (1994) reported electron microprobe geothermobarometry from rocks in the SIP. Both papers identified the following three stages of metamorphism:

- M1- Contact metamorphism preserved in xenolithic rafts of paragneiss within metamorphosed gabbro. This assemblage is marked by orthopyroxene, garnet, plagioclase, K-feldspar, hercynite spinel, ilmenite, and sometimes cordierite within the xenoliths. This phase is associated with pure CO₂ inclusions, and took place at temperatures of 750- 950°C and pressures of 5 kbar.
- M2- Regional metamorphism at some point between 850 Ma and 550 Ma. This phase of metamorphism is responsible for the formation of a foliation in both the paragneiss and the gabbros. This metamorphic event is marked by the presence of clinopyroxene, orthopyroxene, plagioclase, ilmenite and sometimes hornblende in the mafic rocks of the SIP, and garnet, sillimanite, orthopyroxene, k-feldspar, quartz, plagioclase, biotite and ilmenite in the paragneisses adjacent to the gabbros. Fluid inclusions with a mixed CO₂-N₂ composition mark this

phase of metamorphism. Temperatures of 700- 750°C and pressures of 5- 7 kbar are estimated.

- M3- Regional metamorphism after 500 Ma. This phase of metamorphism is limited to narrow ductile shear zones, and is marked by the overgrowth of new minerals on the M2 assemblage. This phase of metamorphism is marked by garnet, clinopyroxene, orthopyroxene, plagioclase, rutile and quartz in the gabbros, and garnet, kyanite, quartz, plagioclase, K-feldspar, biotite and rutile in the paragneisses. This event is linked to pure N₂ inclusions in late- stage garnet. Temperatures of 650- 700°C and pressures of 8- 10 kbar are estimated.

This metamorphic scheme, although intricate and based on a significant amount of electron microprobe data, does not fit with the conclusions reached during this study.

The previous metamorphic work is based on two assumptions that can be questioned:

- Metamorphic events M2 and M3 are based on the correlation of metamorphism in the gabbros with metamorphism in the paragneisses. If the paragneiss is much older than the gabbros, the metamorphic assemblages present in the rock may be related to earlier metamorphic events.
- The time gaps postulated in this scheme are based on flawed geochronology. If only two significant metamorphic events have affected the SIP, it is possible that either M1 and M2, or M2 and M3, represent different parts of the same metamorphic event.

6.3. Results from U-Pb age dating of Øksfjord paragneiss

The paragneissic body sampled by Elvevold et al. (1994) and used to defined the metamorphic phases M2 and M3 is found as a long ribbon of gneissic material running through the Øksfjord peninsula (Figure 6.1). This paragneiss is a highly foliated, garnetiferous rock in outcrop (Fig. 6.2).

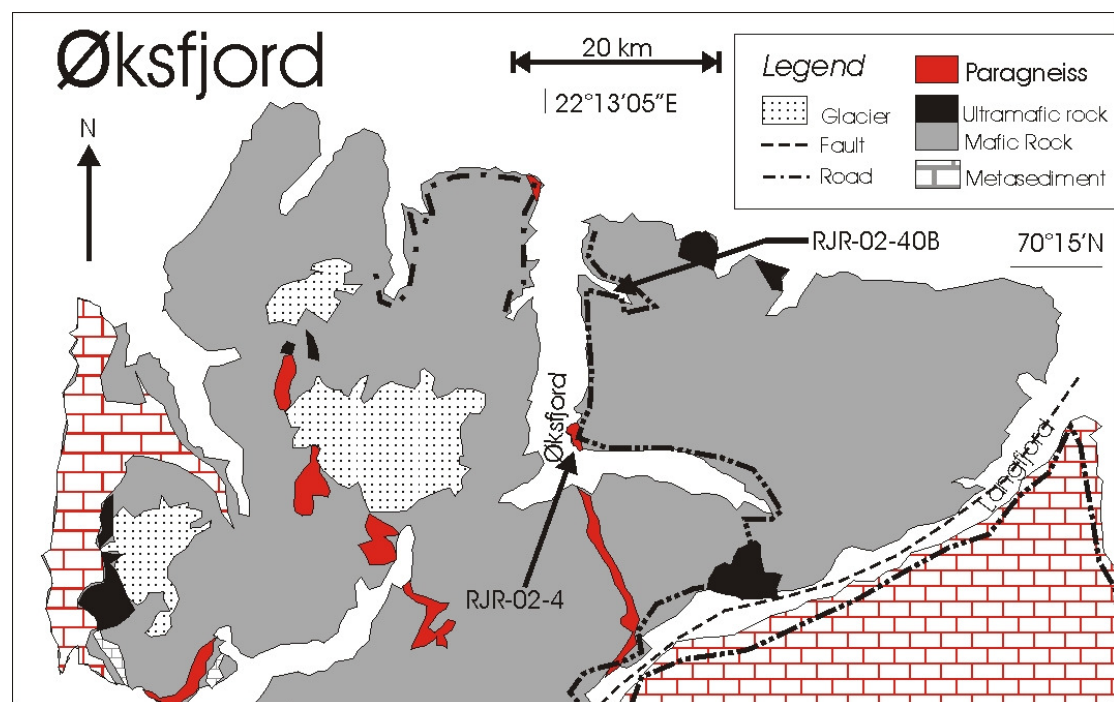


Figure 6.1- Map of the Øksfjord peninsula, showing the paragneiss sampled by Elvevold et al. (1994), and the position of the sample taken for this study.



Figure 6.2- Paragneiss, as developed at location RJR-02-4. The paragneiss hosts mafic dykes which has experienced post-emplacement strain, resulting in a boudinaged structure.

One sample of the paragneiss, RJR-02-4, was analysed for U-Pb ID-TIMS dating (Table 6.1). Five fractions of zircon and two monazites were analysed. The zircons are extremely discordant, with a lower intercept of 592 ± 190 Ma and an upper intercept of 1459 ± 360 Ma (Fig. 6.3). The lower intercept is similar in age to the SIP, and the upper intercept is similar to the proposed age for Klubben Quartzite (Kirkland & Daly, 2003), but the age is extremely imprecise. However, more importantly for the matter at hand, two monazites retrieved from the paragneiss are concordant. These two monazites plot at 641 and 635 Ma, and are considered to represent an age of metamorphic mineral growth in the rock.

Table 6.1: U/Pb analyses of Øksfjord paragneiss

						²⁰⁶ Pb/ ²⁰⁴ Pb ^c	²⁰⁷ Pb/ ²³⁵ Pb ^d	²⁰⁶ Pb/ ²³⁸ Pb ^d	rho	²⁰⁷ Pb/ ²⁰⁶ Pb ^d	²⁰⁷ Pb/ ²⁰⁶ Pb ^e	
No.	Type	Weight [ug]	U [ppm]	Th/U ^a	Pb com ^b [pg]							
RJR02-4B	A	1 small brown zircon	1	637	3.56	0.9	7100	1.6416±0.00760	0.15007±0.0068	0.890	0.07934±0.00017	901.3
	B	Several small clear zircons	10	416	0.27	3.6	10459	1.5408±0.00365	0.14463±0.0003	0.936	0.07726±0.00006	870.8
	C	Several small brown zircons	3	249	0.53	3.5	1996	1.5359±0.0075	0.14716±0.00062	0.890	0.0757±0.00017	885.0
	D	Euhedral brown fragments	6	279	0.20	27.4	493	1.2339±0.0097	0.12413±0.00058	0.550	0.07209±0.00048	754.3
	E	Elongated fragments	2	406	0.37	0.9	7393	1.2493±0.0053	0.12672±0.00049	0.930	0.0715±0.00011	769.1
	M1	Monazite	1	8958	7.63	49.5	1191	0.8697±0.004	0.1035±0.00044	0.760	0.06094±0.00019	634.9
	M2	Monazite	1	781	6.66	7.3	719	0.8797±0.0081	0.1045±0.00041	0.680	0.06106±0.00041	640.7

Footnotes

- a** Model value calculated from 208Pb/206Pb ratio and the age of the sample
- b** Total common lead, including analytical blank and initial common lead in the sample
- c** Corrected for spike contribution and fractionation
- d** Corrected for spike contribution, fractionation, blank and initial common lead (as calculated from Stacey and Kramers, 1975), errors reported at 2σ
- e** 2σ absolute errors reported in Ma

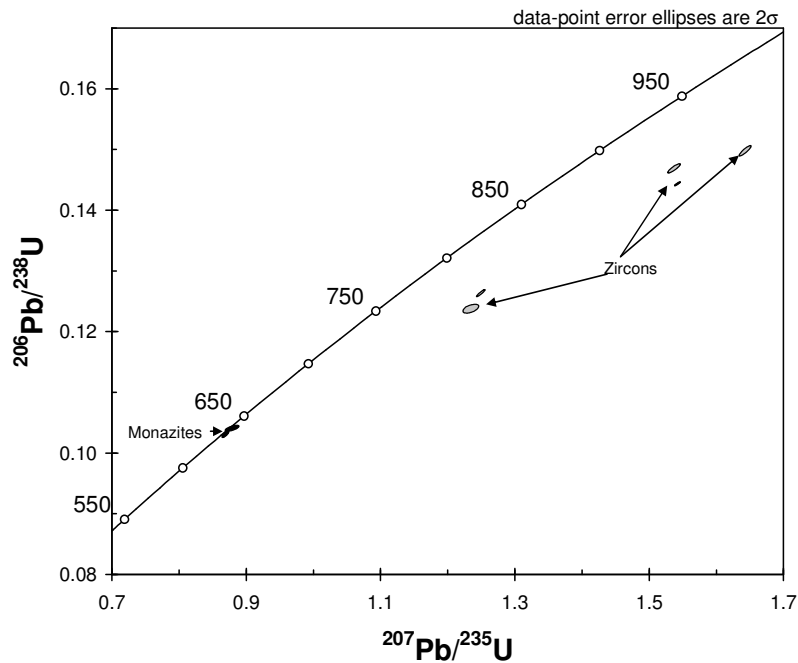


Figure 6.3- Concordia plot for zircons and monazites from paragneiss RJR-02-4.

6.4. Discussion

6.4.1. Metamorphism of paragneiss RJR-02-04

Paragneiss RJR-02-04 clearly records metamorphism that is much older than the emplacement of the SIP. Two monazites plot concordantly at 630- 640 Ma, similar to ages derived from granitoids on Porsanger (Corfu et al., in press), in a rock that is at least Mid-Proterozoic in age. This leads to two conclusions:

- The paragneiss has experienced metamorphism prior to the emplacement of the SIP
- Metamorphism and magmatism accompanying the SIP, and subsequent metamorphic events, has not been sufficient to reset the age given by the monazites in the paragneiss.

Under the metamorphic scheme postulated by Elvevold et al. (1994), the oldest metamorphic assemblage in the paragneiss was deemed to be related to the intrusion of the SIP. The new monazite data makes it clear that this was not the case. Therefore, it is inappropriate to use metamorphic petrology from the paragneiss in conjunction with metamorphic data from the metamorphosed mafic rocks of the SIP, as metamorphic events cannot be reliably correlated between the older paragneiss and younger gabbros.

6.4.2. Reappraising the M2 metamorphic event

If the geothermobarometry from Elvevold et al. (1994) is re-evaluated without including any data from the Øksfjord paragneisses, then a radically different view of the metamorphic history of the SIP emerges. Of the three metamorphic events recognised by Elvevold et al (1994), the event most dependent on data from the paragneiss is the M2 event. A garnet-orthopyroxene thermometer (Powell and Holland, 1988) and a garnet- Al_2SiO_5 - plagioclase- quartz barometer (Hodges & Crowley, 1985) from the paragneisses, supported by two-pyroxene thermometry from the gabbros (Powell and Holland, 1988), originally defined M2 metamorphic pressures and temperatures of 5-7 kbar and 700- 750°C, respectively. Excluding the paragneiss data leaves only a temperature estimate of 700- 750°C for the M2 event. This temperature range is similar to the temperature range postulated for the later M3 event (Elvevold et al., 1994), and removes the necessity to consider the M2 event as a metamorphic event distinct from either M1 or M3.

6.4.3. A metamorphic model for the SIP

The U-Pb systematics preserved in the zircons of the Breivikbotn Alkaline Complex (Chapter 4) indicate that only one episode of metamorphism postdates the emplacement of the SIP in the Kalak Nappes. As the identification of M1, a contact metamorphic event, is constrained to data obtained from xenolithic rafts in the gabbros of the SIP, it seems likely that the mineral assemblages M2 and M3, as preserved in the igneous rocks of the SIP and identified by Elvevold et al (1994), represent the same metamorphic event. Thus, a model for the metamorphic history of the SIP that best fits the available data is the following:

- 1) The sediments forming the Kalak Nappe Complex were originally deposited at or before 1800 Ma (Kirkland & Daly, 2003; Corfu et al., in press), and deformed and metamorphosed at least once, at 630- 640 Ma (M0- P, T conditions unknown, but sufficient to allow the growth of monazite).
- 2) The igneous rocks of the SIP were emplaced during extensional rifting at 560- 570 Ma, at pressures of 5.5- 7kbar (20-30 km depth). Contact metamorphism accompanied the emplacement of the SIP plutons, with local temperatures between 500 and 900°C (M1).
- 3) Subsequent to the emplacement of the SIP, the Kalak Nappes were involved in the Scandian (420 Ma) phase of the Caledonian Orogeny. During this event, the rocks experienced a pressure increase, and peak metamorphic conditions are estimated at 650- 750°C and 8-10 kbar (M2).

This new model needs to be evaluated thoroughly by careful petrological investigation, but is considered a more reasonable model of the evolution of the SIP than the previous three-stage model presented by Elvevold et al, (1994).

Chapter 7- Conclusions

The Seiland Igneous Province is hosted as a discrete terrane within the northernmost part of the Caledonian orogenic belt. The Province consists of numerous mafic and ultramafic plutons emplaced into a sedimentary succession indicative of a continental setting. The plutons are relatively small in area, but are numerous, with more than ten discrete mafic plutons and five large ultramafic bodies having been detailed in the literature. Accompanying this mafic magmatism is a significant volume of intermediate monzonitic and dioritic rock (10% of the total exposed igneous rock), as well as numerous nepheline syenite and carbonatitic intrusive material. This alkaline rock is present in both discrete complexes and as distributed dykes, and is generally crosscutting in its relationship with the mafic plutons that host this magmatism. This collection of mafic and ultramafic plutons and their associated intermediate and alkaline intrusions has been interpreted in several different ways in the past.

Previous geochronological studies have yielded inconsistent results for different plutons within the Seiland Igneous Province. This study reports ID-TIMS U-Pb analyses on zircon and lesser monazite from a variety of different igneous rocks across the Province. Whereas previous studies had produced a range of ages between 420 Ma and 830 Ma, this study has shown that primary igneous zircons from not only the mafic plutons but also monzonitic, granitic and alkaline intrusions from Øksfjord and Sørøy are all of similar ages. These analyses, from rocks previously considered to be of different ages (age spans of up to 300 m.y.), indicate that the bulk of the Seiland magmatism took place between 560 Ma and 570 Ma. These data contrast with the previous isotopic work. For instance, Sm-Nd analysis of the Hasvik Gabbro produced

an age of 700 Ma, which can be contrasted with the age of 562 Ma obtained from zircons from the intrusion. The age data thus indicate that only one magmatic episode is represented in the rocks of the Seiland Igneous Province, invalidating previous models involving multiple rifting events over a period of 300 m.y.

Furthermore, the close spatial and temporal relationship between the mafic magmatism and the alkaline magmatism that generally crosscut the mafic plutons indicates that the most likely setting for the emplacement of the Seiland Igneous Province is an extensional setting, possibly in an intracontinental rift or in a back-arc setting. It is also apparent from the age of the magmatism and the age of the sediments into which the magma was emplaced that the Seiland terrane need not be para-autochthonous to the Baltican mainland, and could easily be allochthonous in origin.

The contemporaneous nature of the magmatism is important to bear in mind when considering the geochemical characteristics of the Seiland Igneous Province. Previous studies had divided the mafic plutons into three different classes on structural, chronological and petrological grounds. This study has shown that this classification is relevant and can be sustained on geochemical grounds. The detailed investigation of several plutons from an evolved high alkali suite of gabbroic intrusions has shown that these plutons are generally enriched in rare earth elements and other trace elements compared to layered intrusions from other areas across the globe, but that geochemically the gabbros are relatively homogenous, far more than would be expected from a suite of plutons undergoing relatively dissimilar fractionation and emplacement processes.

This chemical homogeneity can be contrasted with the isotopic heterogeneity of the mafic plutons. The rocks yield Lu-Hf and Sm-Nd isotopic values that range from mantle to crustal values, with those plutons insulated from the surrounding country rock showing the most primitive isotopic values (the ϵ_{Hf} and ϵ_{Nd} values for the gabbroic rocks range from +8 to -6 and from +4 to -4, respectively). The most primitive isotopic values are similar to those obtained from the carbonatites and nepheline syenites associated with the igneous province, which indicates that a similar mantle source gave rise to the magmas which were subsequently emplaced as the Seiland Igneous Province. The spread of isotopic values amongst the mafic plutons is considered indicative of crustal contamination.

It is difficult to reconcile the isotopic and geochemical characteristics of the rocks. The homogeneous trace element content of the different mafic rocks most likely indicates a relatively homogeneous mantle source for the original magmas of the province. However, processes of assimilation and crustal contamination have contributed to the current composition of the igneous rocks, both during the ascent of the magmas and during their emplacement. Such contamination is not visible in the trace element profiles, which may indicate that the amount of contaminant involved was relatively small, but of significantly different isotopic composition in order to produce the isotopic variation observed.

It is also apparent in dealing with the isotopic data that many of the monzonitic and dioritic bodies in the Seiland Igneous Province are not derived from melted silicic crustal material, since these rocks report isotopic values of clearly mantle provenance. Since it is unlikely that large amounts of magma of intermediate composition could be

produced by fractionation of a mafic melt, it has tentatively been concluded that the intermediate rocks of the province have been formed by the melting of pre-existing crustal mafic material. This hypothesis requires significant petrological and geochemical investigation before it can be confirmed.

The new geochronology also poses numerous problems for the metamorphic framework previously proposed for the Seiland Igneous Province. There is direct evidence from the zircons that the alkaline rocks were subjected to metamorphism at 420 Ma during the Caledonian Orogeny, and indirect evidence from the mafic rocks that they also underwent metamorphism at this time. However, there is no evidence for metamorphic activity in the period between 570 Ma and 420 Ma, and there are monazites in gneissic rocks hosted within mafic rocks of Seiland age that preserve an age of 640 Ma. This leads to the conclusion that only one metamorphic event affected the Seiland terrane after the emplacement of the Seiland magmas, and invalidates much of the previous geothermobarometry, which was based on coeval metamorphism in both the gneissic and mafic rocks in the terrane. Therefore, a new metamorphic investigation is required to incorporate the new age scheme for the Seiland magmatism with the metamorphic development of the areas.

This study has provided a large amount of new information on the igneous rocks of the Seiland Igneous Province. These data have led to a model for the evolution of the Seiland Province in which a number of heavily modified and contaminated mafic magmas derived from the mantle were emplaced into the lower continental crust of the Seiland nappe between 560 and 570 Ma. This magmatism was accompanied by the injection of volatile-rich alkaline magmas into the same area of the crust, and the

melting of mafic rock emplaced earlier. This magmatic event is considered to have occurred in an extensional stress regime, possibly during intracontinental rifting or back-arc spreading.

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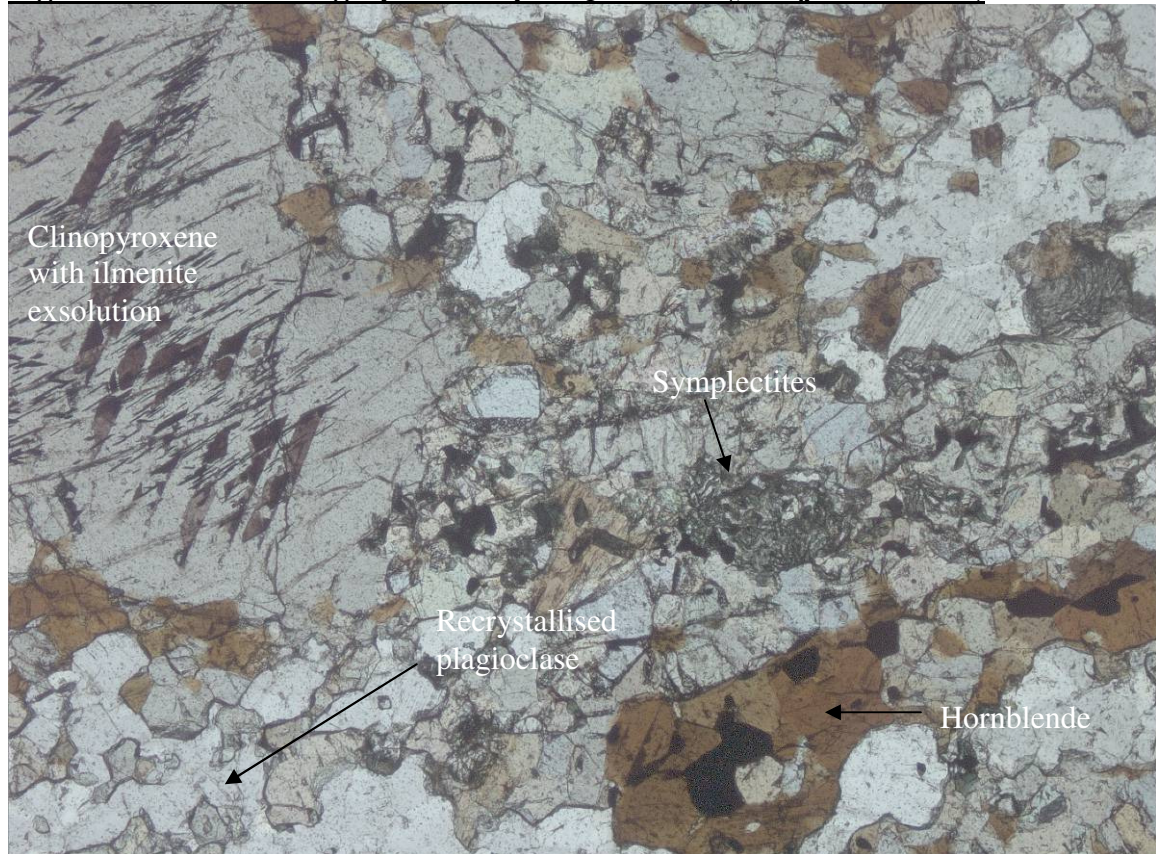
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Appendix A:

**Photomicrographs of samples
used in this study.
Listed by sample number.
Details on Table 2.1.**

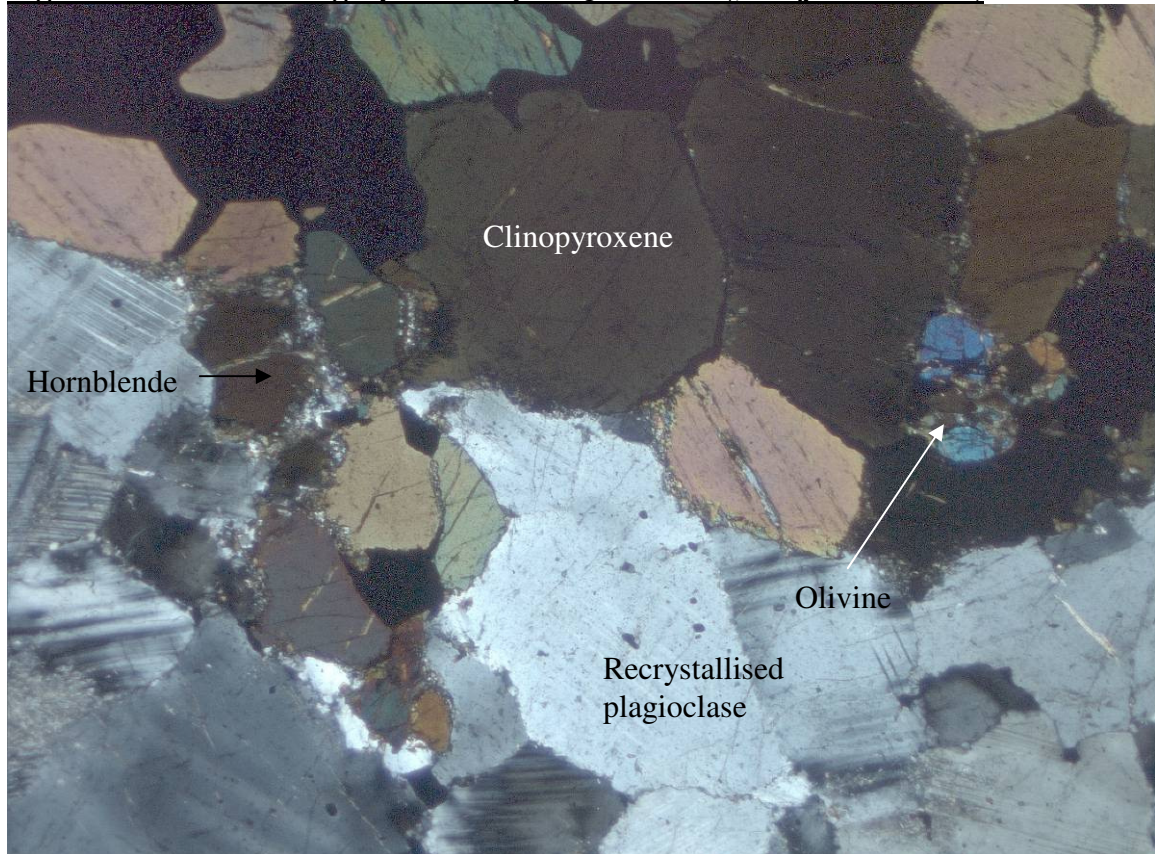
Figure A.1- Photomicrograph of sample RJR-02-2A (Øksfjord Gabbro)



(5x magnification- field of view 5 mm. Plane- polarized light)

This rock consists primarily of brown metamorphic hornblende (40%) and small recrystallised grains of plagioclase (30%). It also consists of large platy crystals of clinopyroxene (30%), with resorbed edges and ilmenite exsolution lamellae along the cleavage planes. The opaques (5%) in this section are predominantly of ilmenite. Also present are numerous symplectic intergrowths of orthopyroxene and green spinel (5%). There are no original igneous textures preserved in this rock.

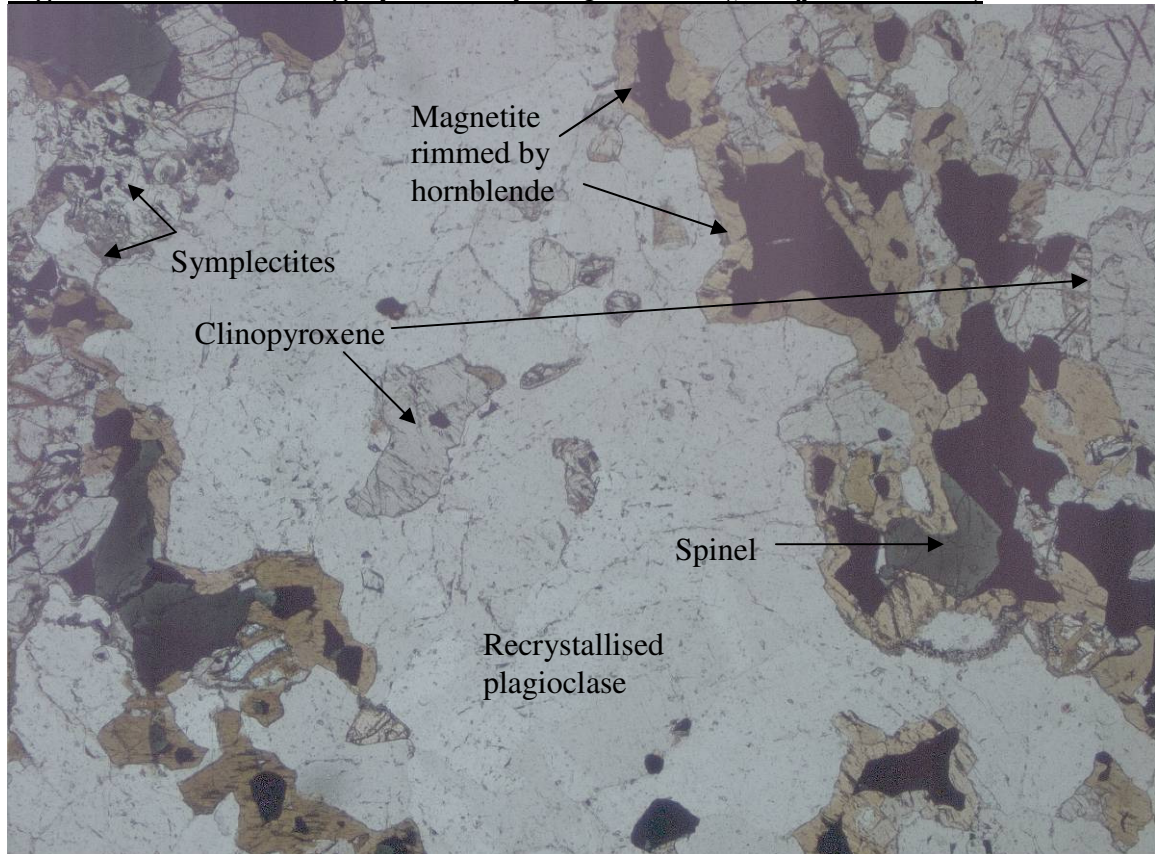
Figure A.2- Photomicrograph of sample RJR-02-3D (Øksfjord Gabbro)



(Field of view 10mm. Cross-polarised light)

This rock consists primarily of clinopyroxene (35%) and metamorphic hornblende (40%), with some plagioclase (20%). Olivine (1%) and orthopyroxene (2%) are minor phases, as are opaques (1%). The plagioclase crystals have been extensively recrystallised, as can be seen in the photomicrograph above, and there is very little of the original cumulus texture remaining.

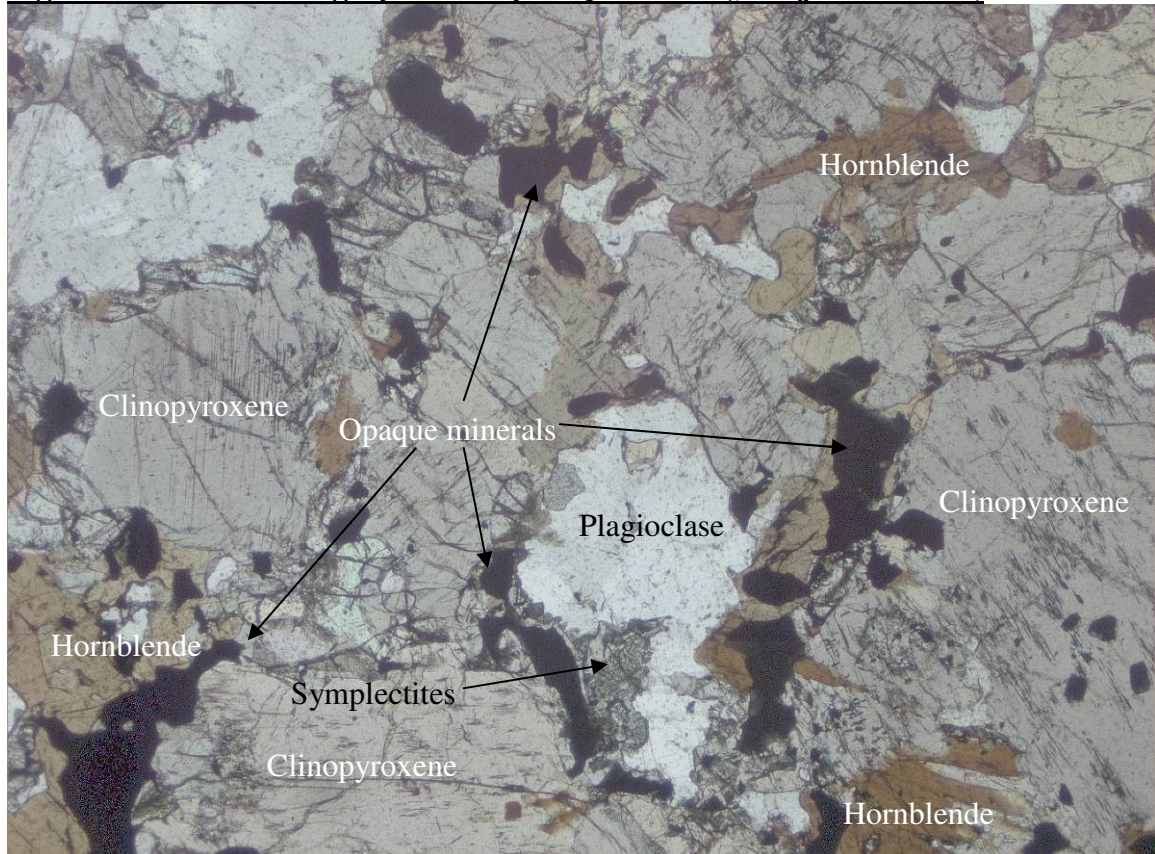
Figure A.3- Photomicrograph of sample RJR-02-6B (Øksfjord Gabbro)



(Field of view 5 mm. Plane-polarised light)

Recrystallised plagioclase (60%) dominates this rock, which shows abundant signs of amphibolite facies metamorphism. Brown metamorphic hornblende (10%) is found as an overgrowth on olivine (5%), clinopyroxene (10%) and opaque minerals (5%). Orthopyroxene-spinel symplectites are present, as is a large amount of green spinel (10%), which may be primary in origin.

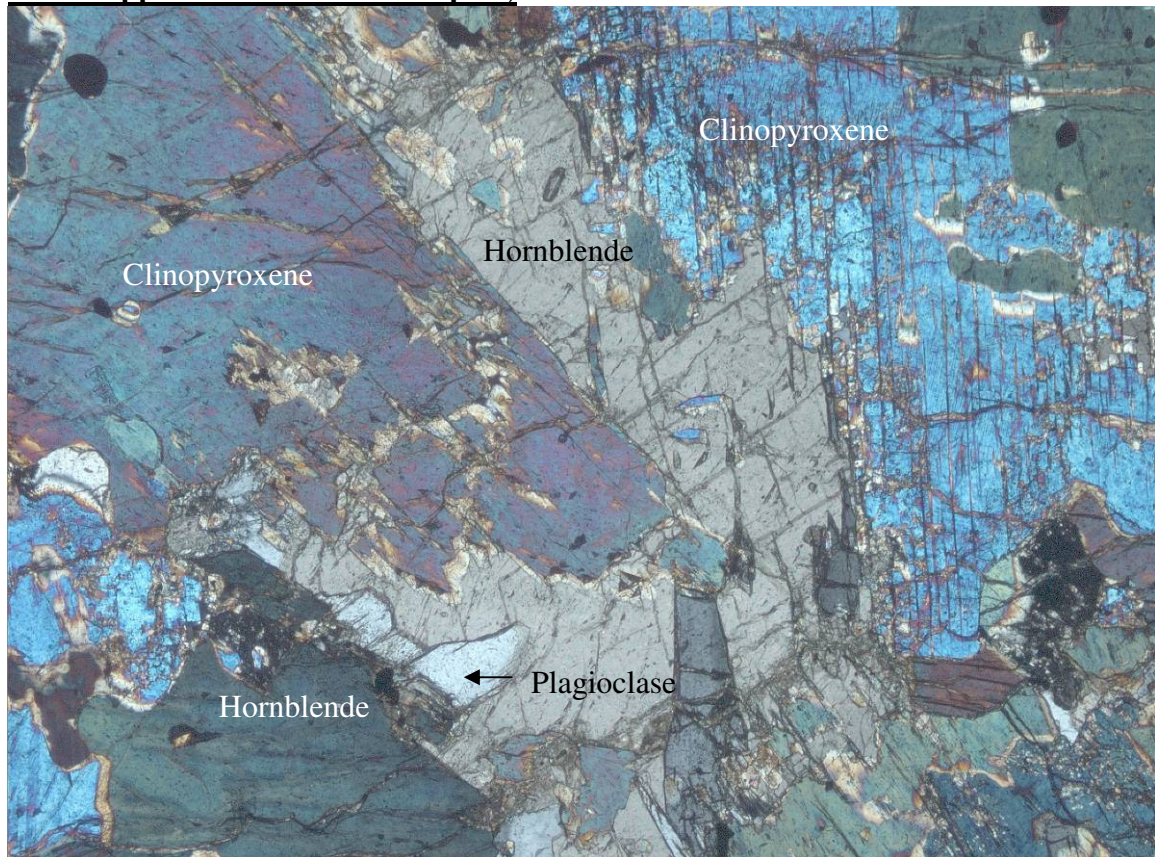
Figure A.4- Photomicrograph of sample RJR-02-6A (Øksfjord Gabbro)



(Field of view 10mm. Plane-polarised light)

This rock consists of large plates of clinopyroxene (30%), recrystallised plagioclase (30%), olivine (5%) and opaque minerals (5%), with brown metamorphic hornblende (30%) occurring as an overgrowth on pyroxene and opaque crystals. Symplectites of spinel and orthopyroxene are also present.

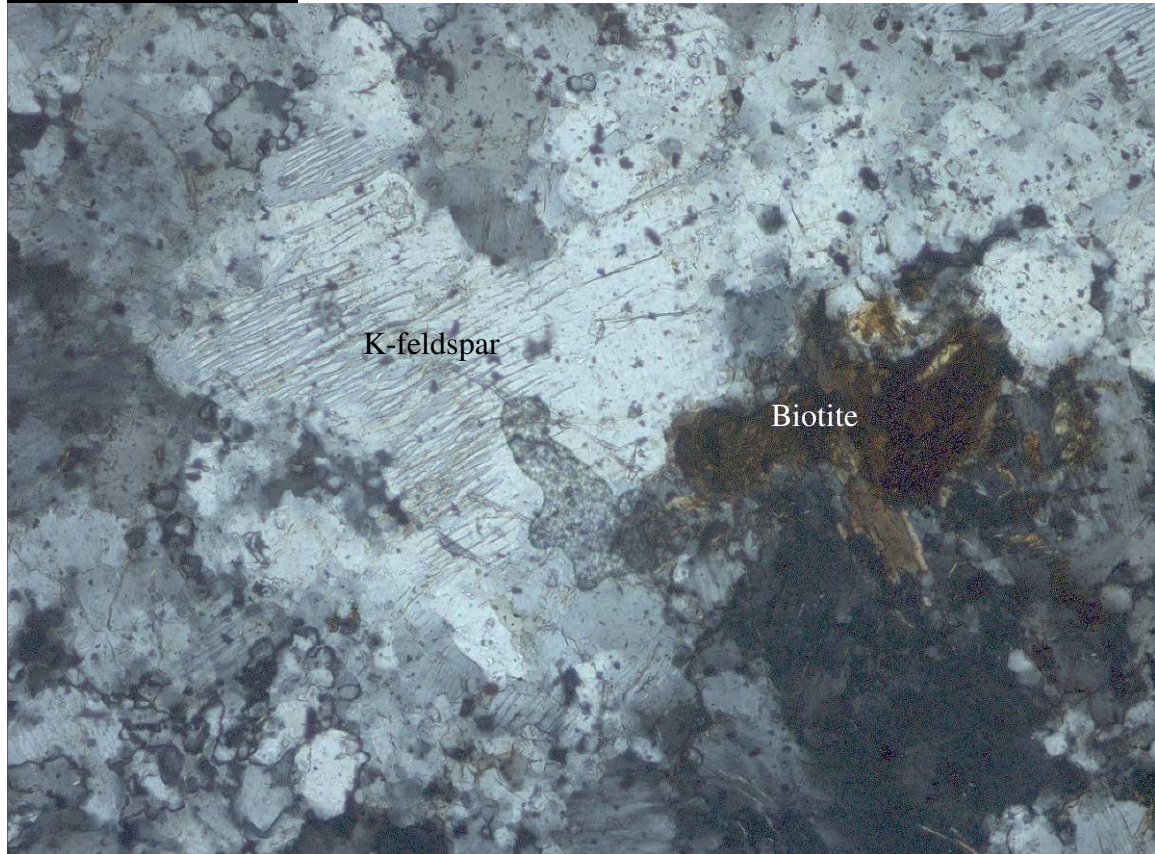
Figure A.5- Photomicrograph of sample RJR-02-7 (Metamorphosed Gabbro from Tappeluft Ultramafic Complex)



(Field of view 10mm. Cross-polarised light)

This gabbro has been exposed to amphibolite facies metamorphism, and consists primarily of brown metamorphic hornblende (60%) and large platy crystals of clinopyroxene (20%). Recrystallised plagioclase is present (10%), as are small amounts of olivine (5%) and opaque minerals (5%). Clinopyroxene and olivine are both resorbed, and little of the original igneous texture is preserved.

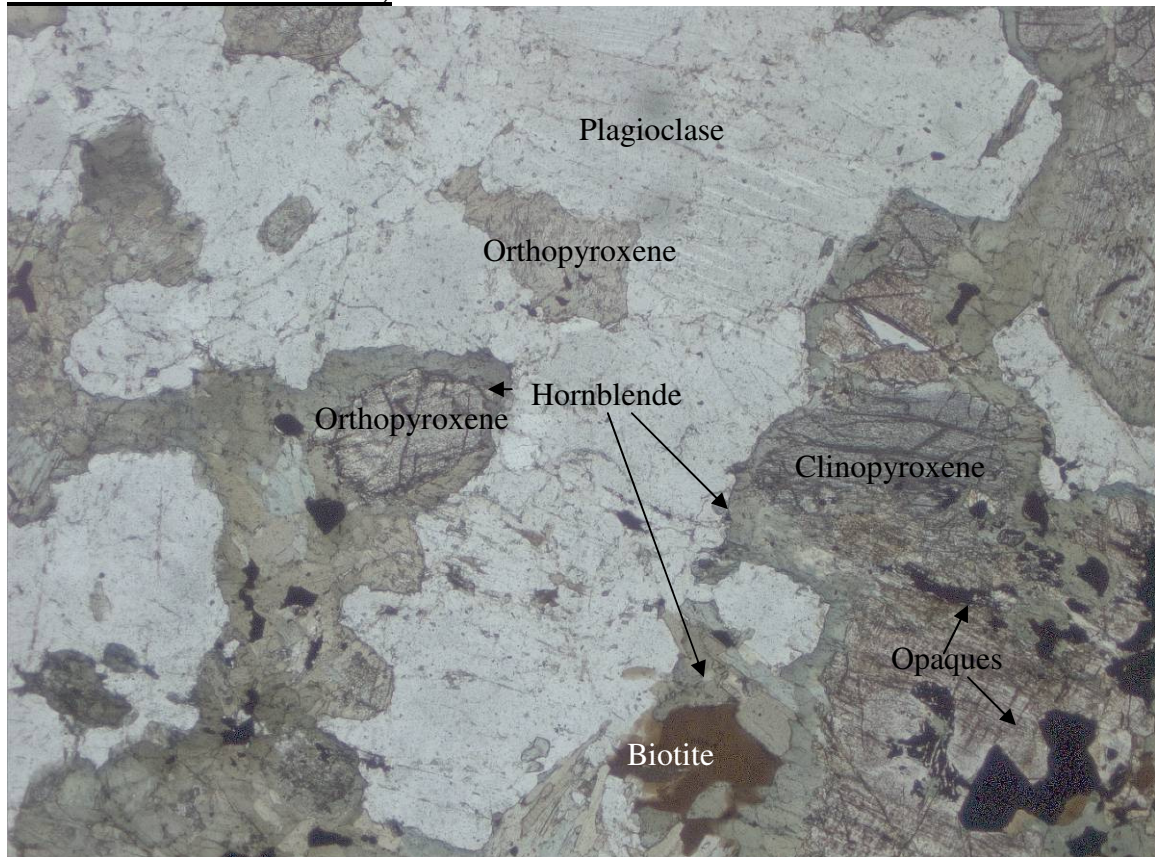
Figure A.6- Photomicrograph of sample RJR-02-8 (Syenite from Tappeluft Ultramafic Complex)



(Field of view 5 mm. Cross-polarised light)

This rock consists primarily of K-feldspar (90%), with subsidiary biotite (5%) and opaque minerals (5%). The rock has a metamorphic fabric defined by thin bands of elongated biotite. The K-feldspar crystals are anhedral and vary greatly in size, and display textures indicative of recrystallisation.

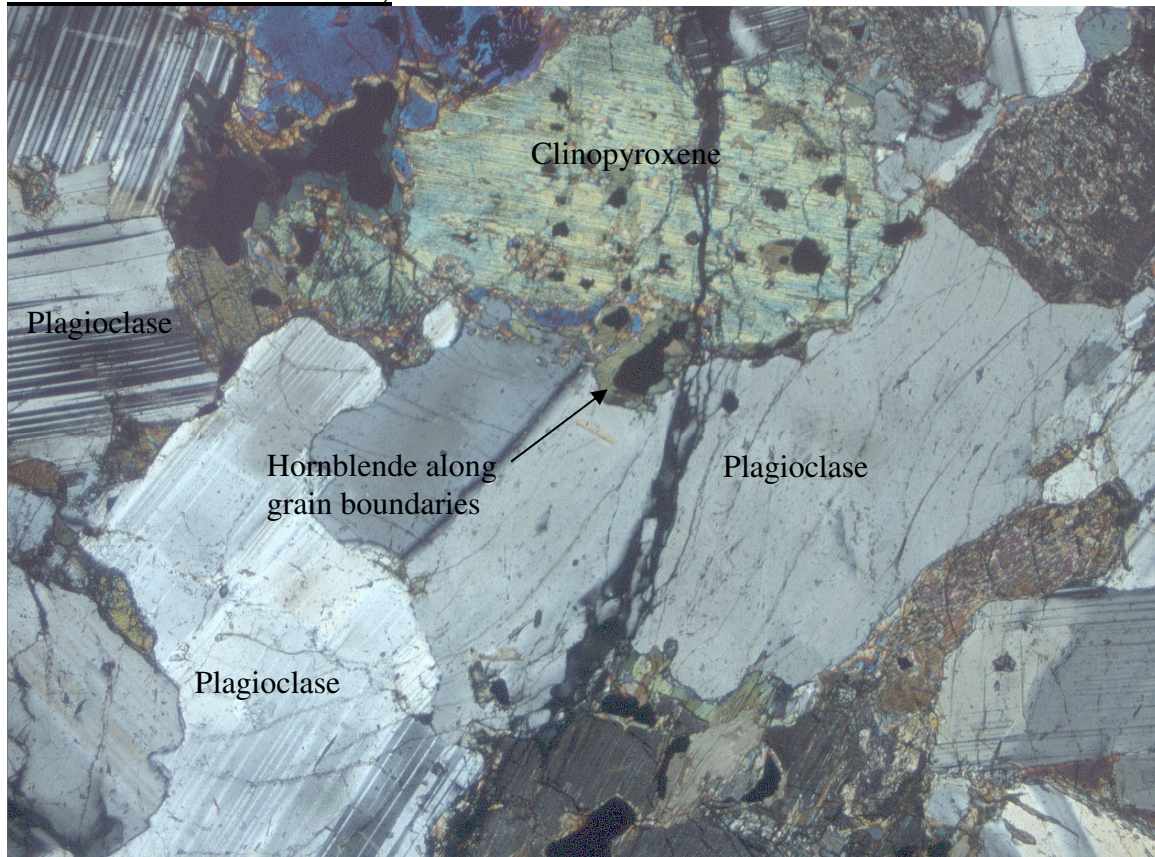
Figure A.7- Photomicrograph of sample RJR-02-30A (Gabbro from the Upper Zone of the Hasvik Gabbro)



(Field of view 10mm. Plane-polarised light)

This rock shows well-developed metamorphic textures, with many igneous crystals showing overgrowths of pale green hornblende. The rock consists of recrystallised plagioclase (40%), hornblende (40%), orthopyroxene (5%), clinopyroxene (5%), biotite (5%) and opaque minerals (5%). Symplectites of magnetite and orthopyroxene are developed in places, and opaque minerals can be seen to be replacing clinopyroxene along cleavage planes. None of the original igneous texture is preserved.

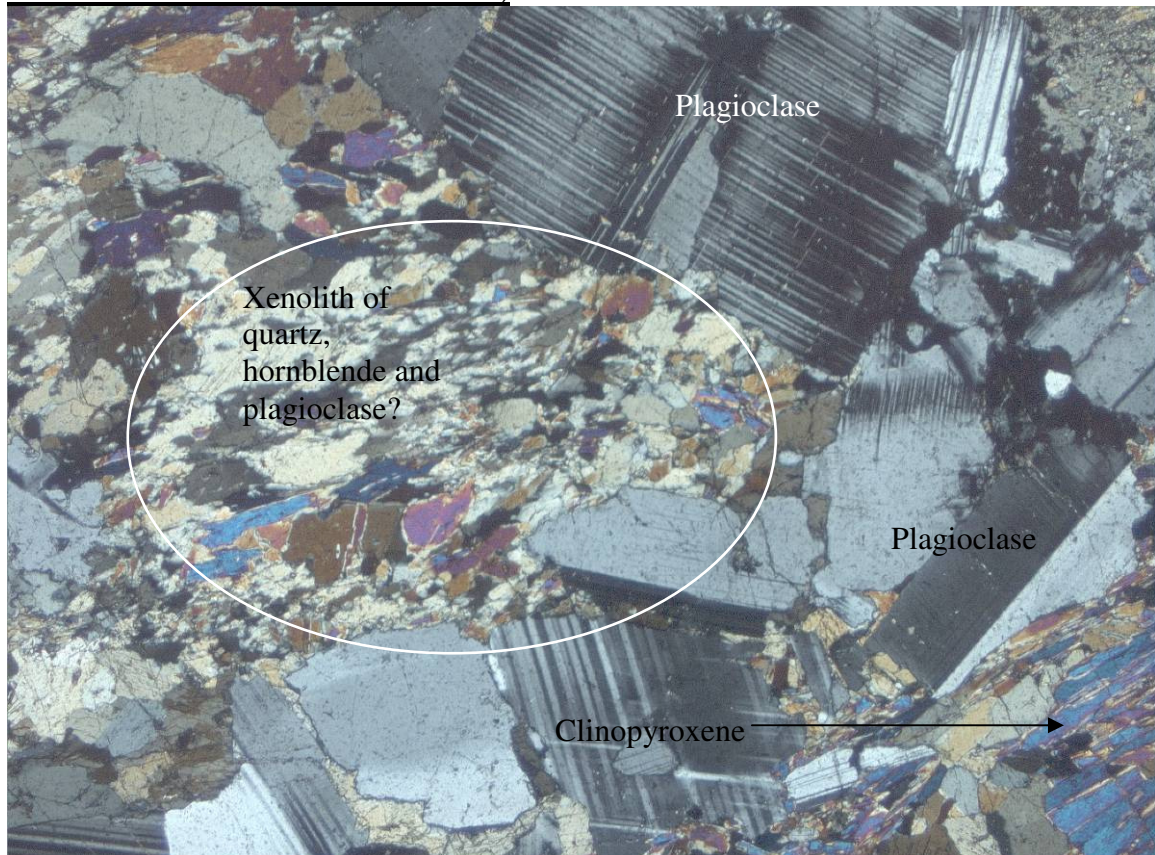
Figure A.8- Photomicrograph of sample RJR-02-30B (Gabbro from the Upper Zone of the Hasvik Gabbro)



(Field of view 10mm. Cross-polarised light)

This rock consists of large laths of recrystallised plagioclase (50%) and large phenocrysts of clinopyroxene (25%) and orthopyroxene (10%), which show alteration to hornblende (10%) and some biotite (2%) along the edges. Opaque minerals (3%) are also present, and symplectic intergrowths of orthopyroxene and magnetite are well developed.

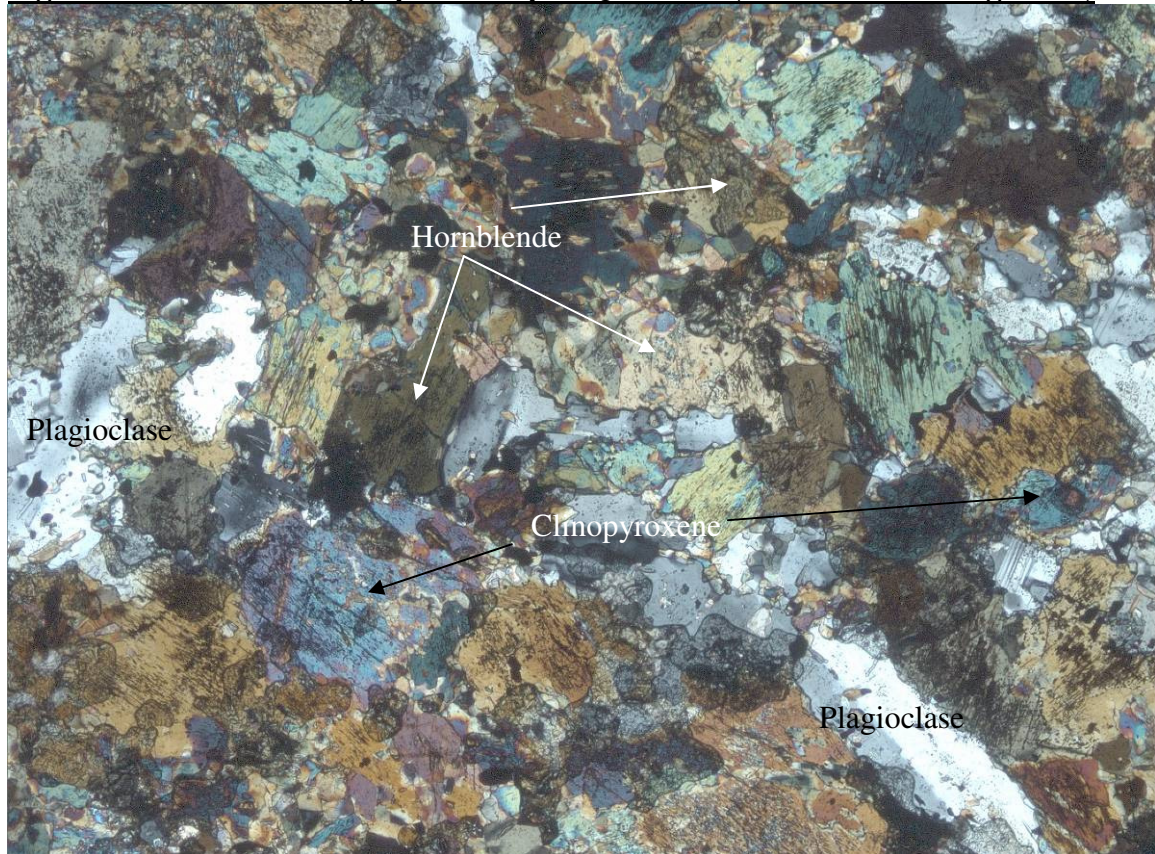
Figure A.9- Photomicrograph of sample RJR-02-30D (Gabbro from the Upper Border Series of the Hasvik Gabbro)



(Field of view 10mm. Cross-polarised light)

This gabbro has an unusual mineralogy, containing a small amount of quartz (3%) along with plagioclase (55%), clinopyroxene (20%), hornblende (20%), and some opaque minerals (2%). It is difficult to distinguish between metamorphic and igneous hornblende in this section, as hornblende is present in intergrown clumps of clinopyroxene and hornblende, with some crystals of hornblende showing good crystal shape. These clumps also contain quartz, and are possibly xenoliths incorporated into the magma as it crystallized. The plagioclase in this section is recrystallised in most places, though some crystals preserve their original cumulus mineralogy.

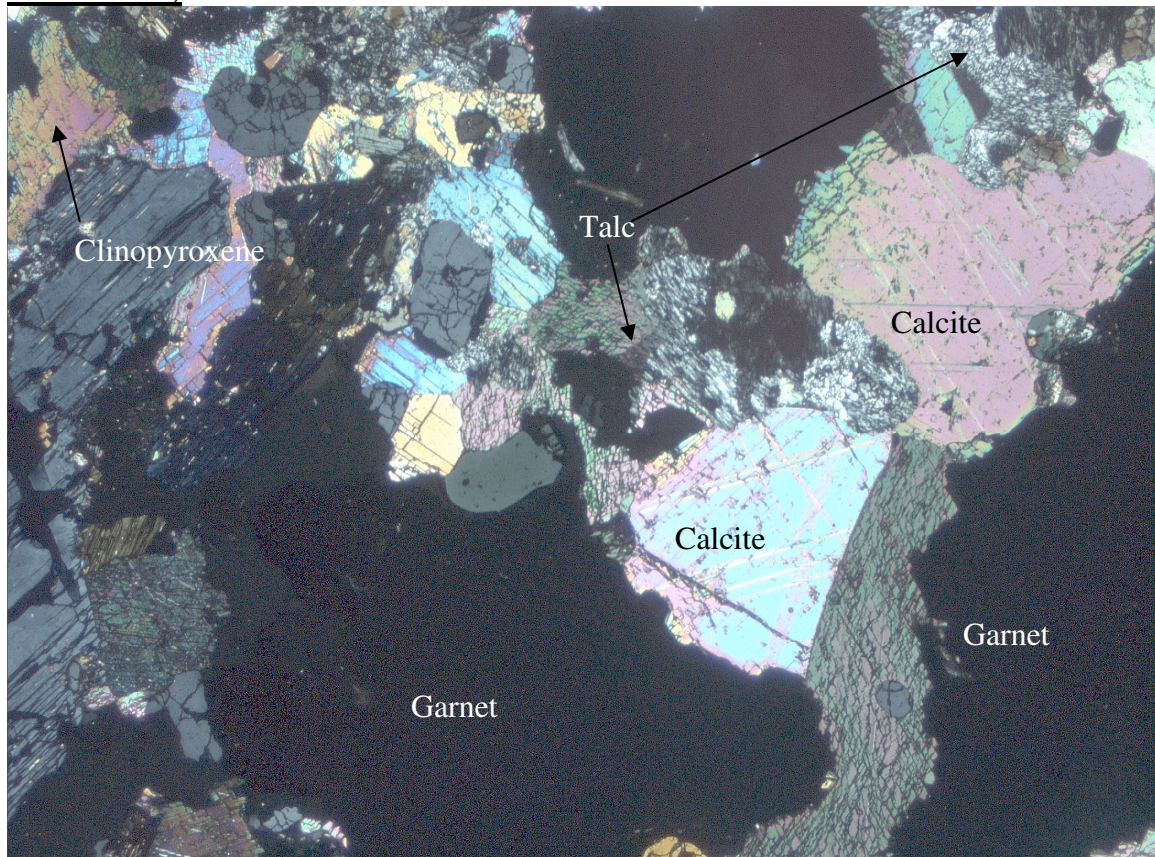
Figure A.10- Photomicrograph of sample RJR-02-31 (Breivikbotn Metagabbro)



(Field of view 10mm. Cross-polarised light)

This gabbro has been undergone metamorphism to amphibolite facies, and consists of metamorphic hornblende (40%), plagioclase (45%), and some remnant crystals of clinopyroxene (5%). Most of the plagioclase has been recrystallised, but the original laths can be seen in places.

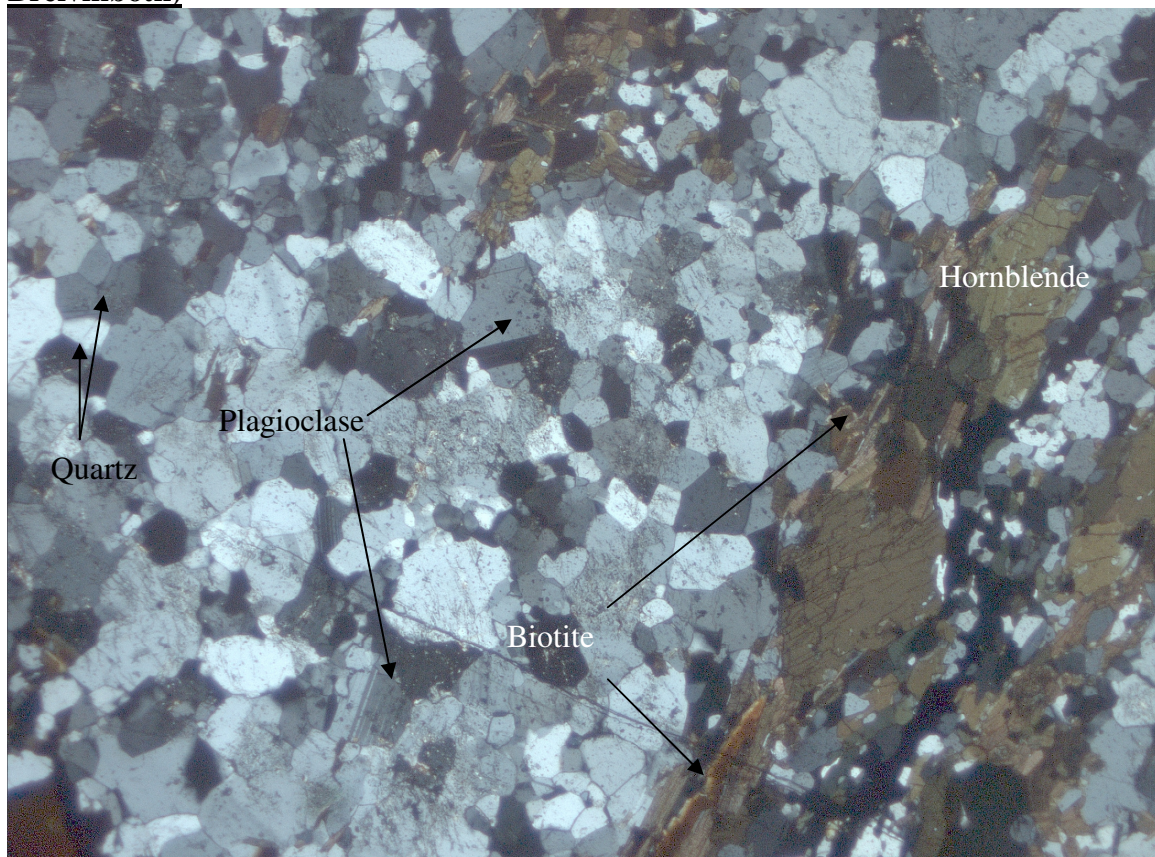
Figure A.11- Photomicrograph of sample RJR-02-34D (Silico-carbonatite from Breivikbotn)



(Field of view 10mm. Cross-polarised light)

This rock consists of calcite (40%), clinopyroxene (20%), talc (20%) and garnet (15%), with some biotite (4%) and apatite (1%). Clinopyroxene occurs as large, early formed, anhedral crystals, showing alteration around the edges. Garnet occurs as large anhedral crystals which include other minerals such as calcite. Talc fills the spaces between the other crystals. The extent of metamorphism is hard to judge in this rock.

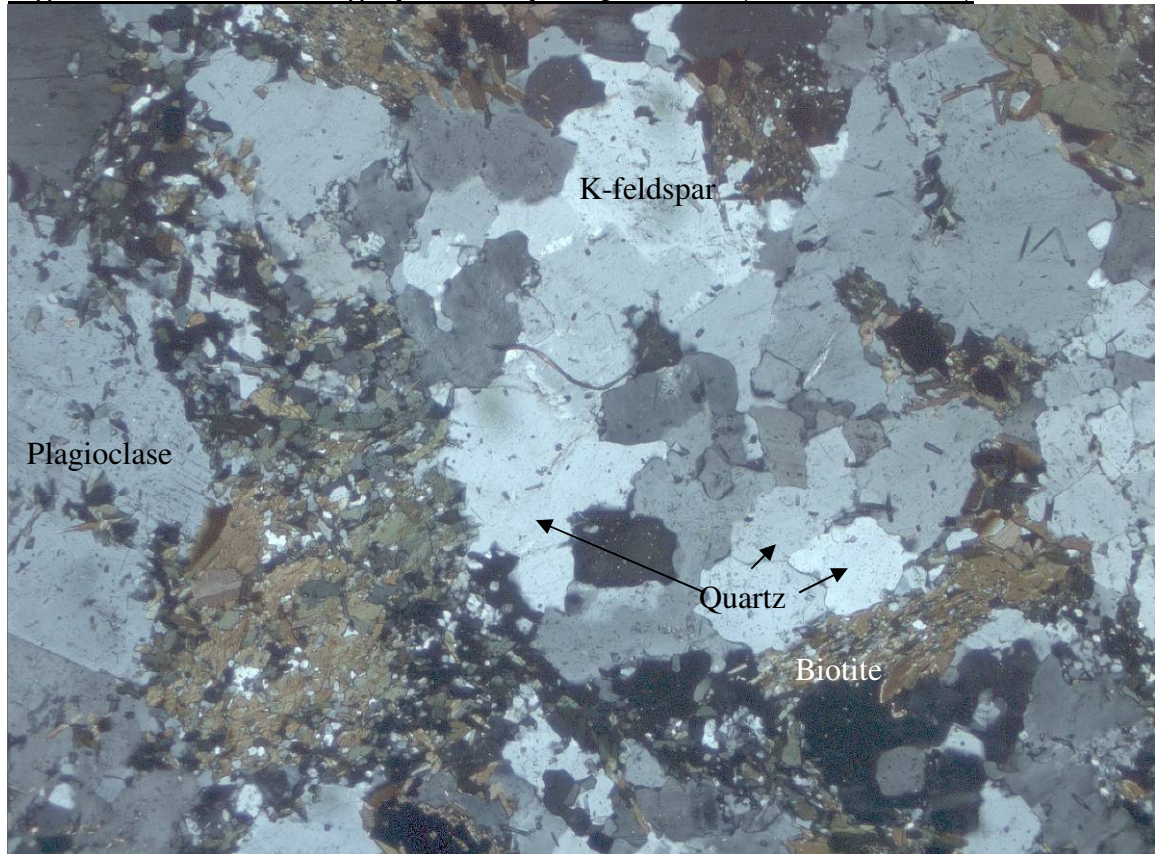
Figure A.12- Photomicrograph of sample RJR-02-35 (Quartz Diorite from Breivikbotn)



(Field of view 10mm. Cross-polarised light)

This rock is marked by a clear foliation created by clumps of elongated biotite crystals. The rock consists of fine recrystallised plagioclase (50%) and quartz (20%), along with biotite (15%), k-feldspar (10%) and hornblende (5%).

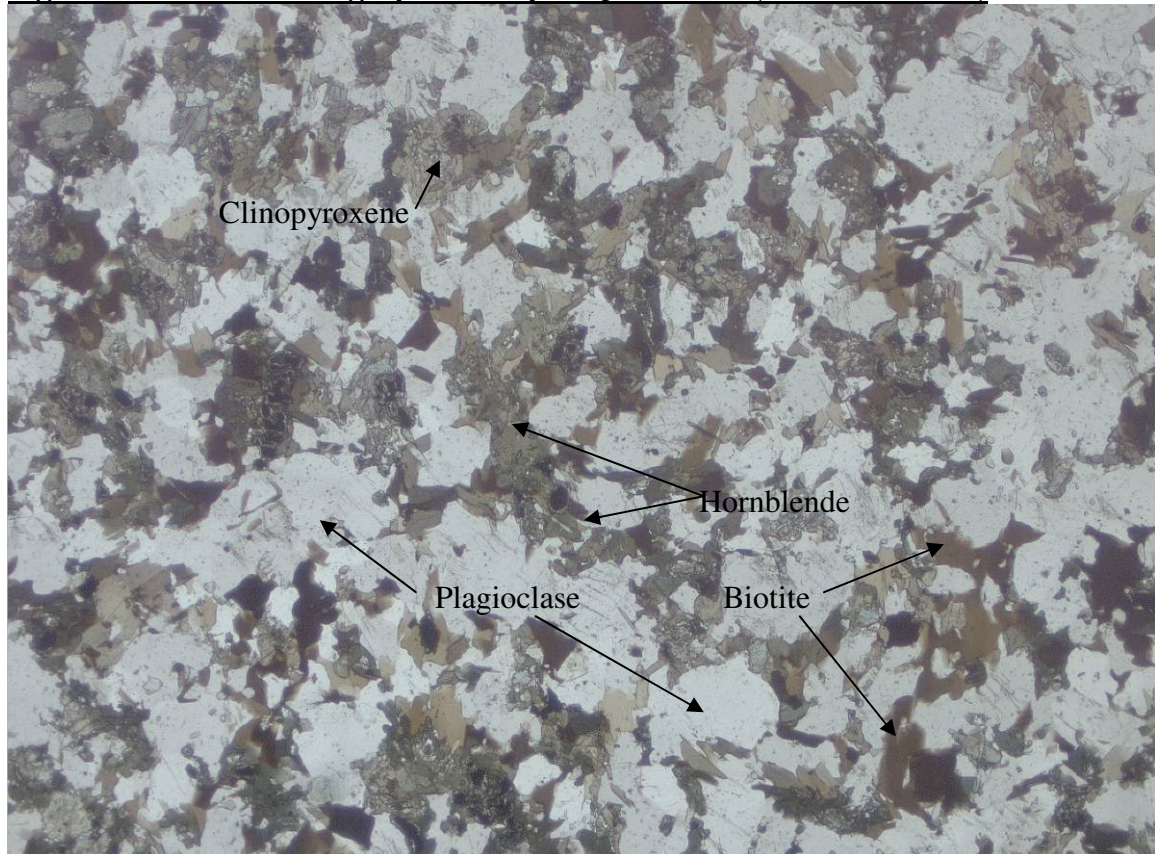
Figure A.13- Photomicrograph of sample RJR-2-37A (Storelv Granite)



(Field of view 10mm. Cross-polarised light)

This rock is a granite containing 50% K-feldspar, 20% hornblende, 20% biotite and 10% quartz. The hornblende and biotite is found in glomeroblastic clumps, and there is no distinct foliation in this rock. K-feldspar and quartz are recrystallised in places, forming fine-grained aggregates along mineral boundaries.

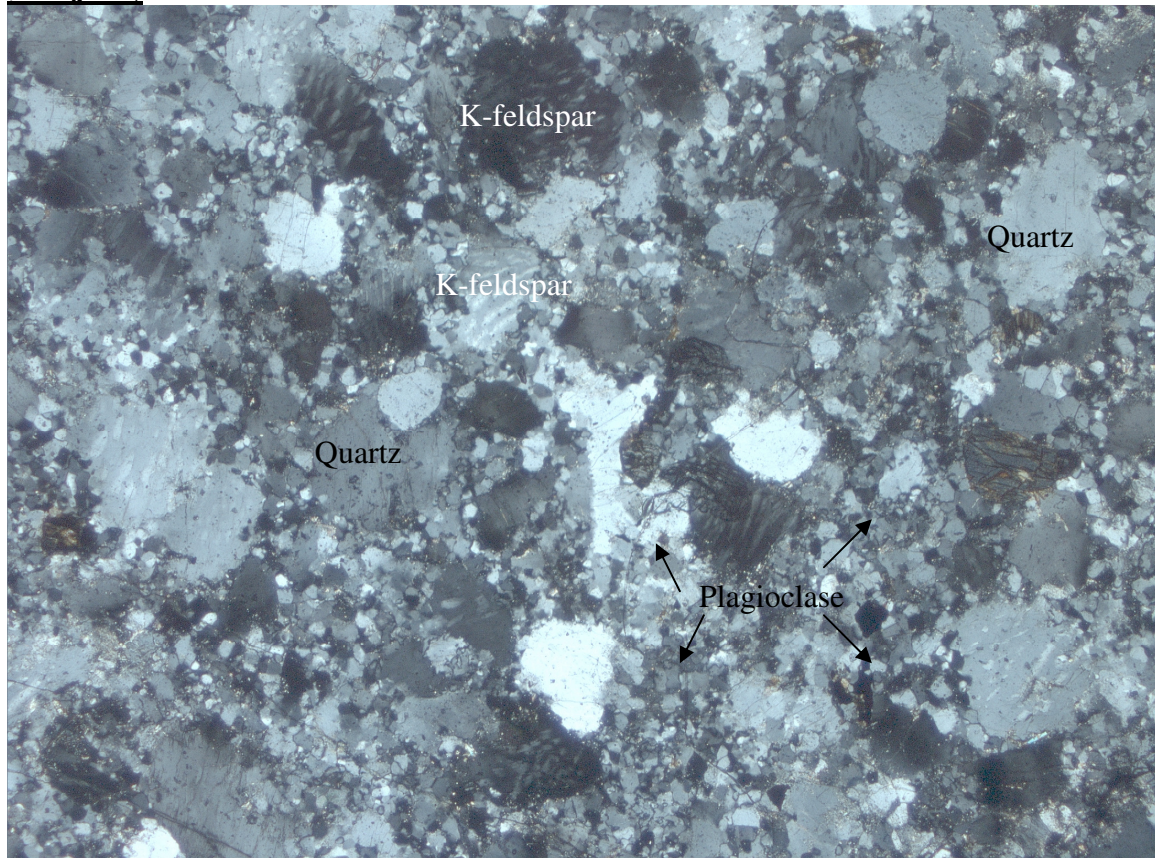
Figure A.14- Photomicrograph of sample RJR-02-37B (Storelv Gabbro)



(Field of view 10mm. Plane-polarised light)

Only a few remnants of orthopyroxene (1%) and clinopyroxene (3%) are preserved in this gabbro. The bulk of the rock consists of recrystallised plagioclase (50%), metamorphic hornblende (20%) and biotite (20%). Small amounts of opaque minerals (5%) and a relatively large amount of apatite (1%) are present in this rock. The embayed pyroxene remnants are generally found within hornblende, and the plagioclase is recrystallised, indicating metamorphism at amphibolite facies.

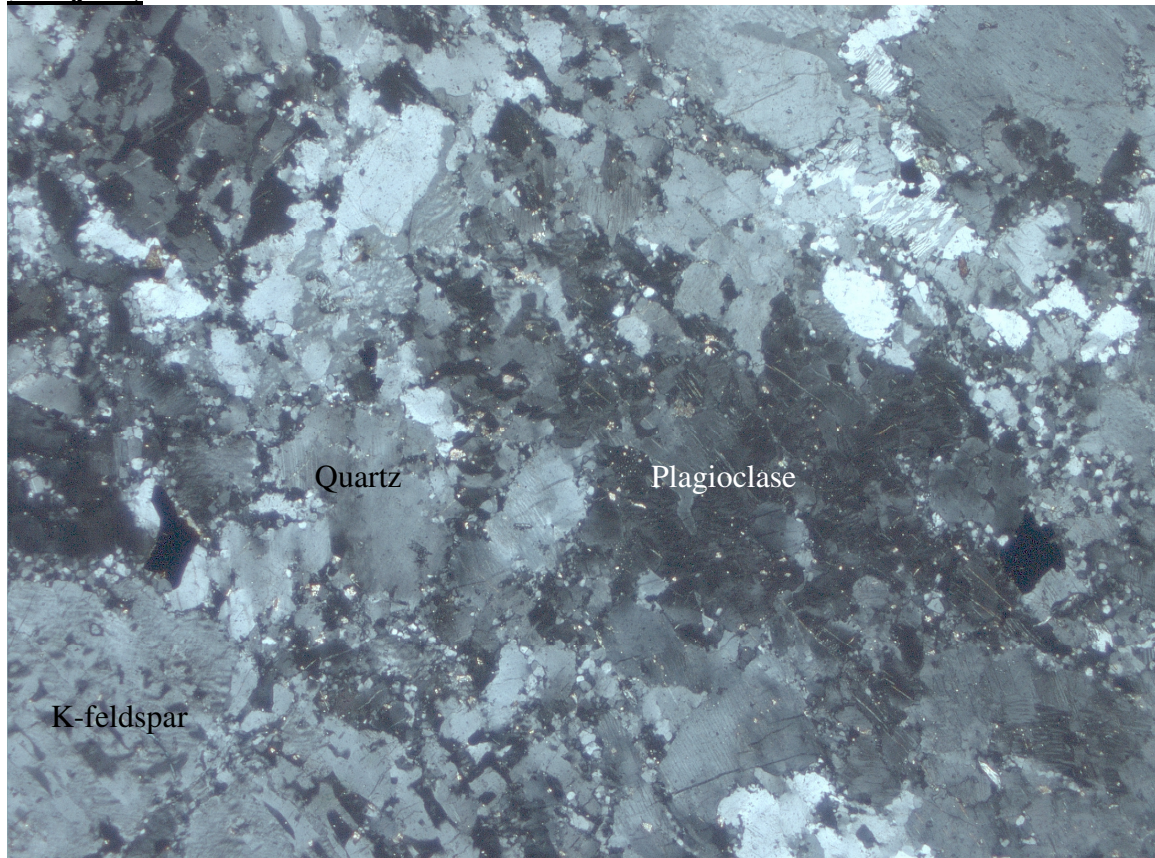
Figure A.15- Photomicrograph of sample RJR-02-40B (Monzonite from Øksfjord)



(Field of view 10mm. Cross-polarised light)

This rock consists of plagioclase (60%), K-feldspar (35%), and some quartz (5%). The feldspars occur as large anhedral phenocrysts in a recrystallised matrix of feldspar and quartz, and little of the original igneous texture is preserved.

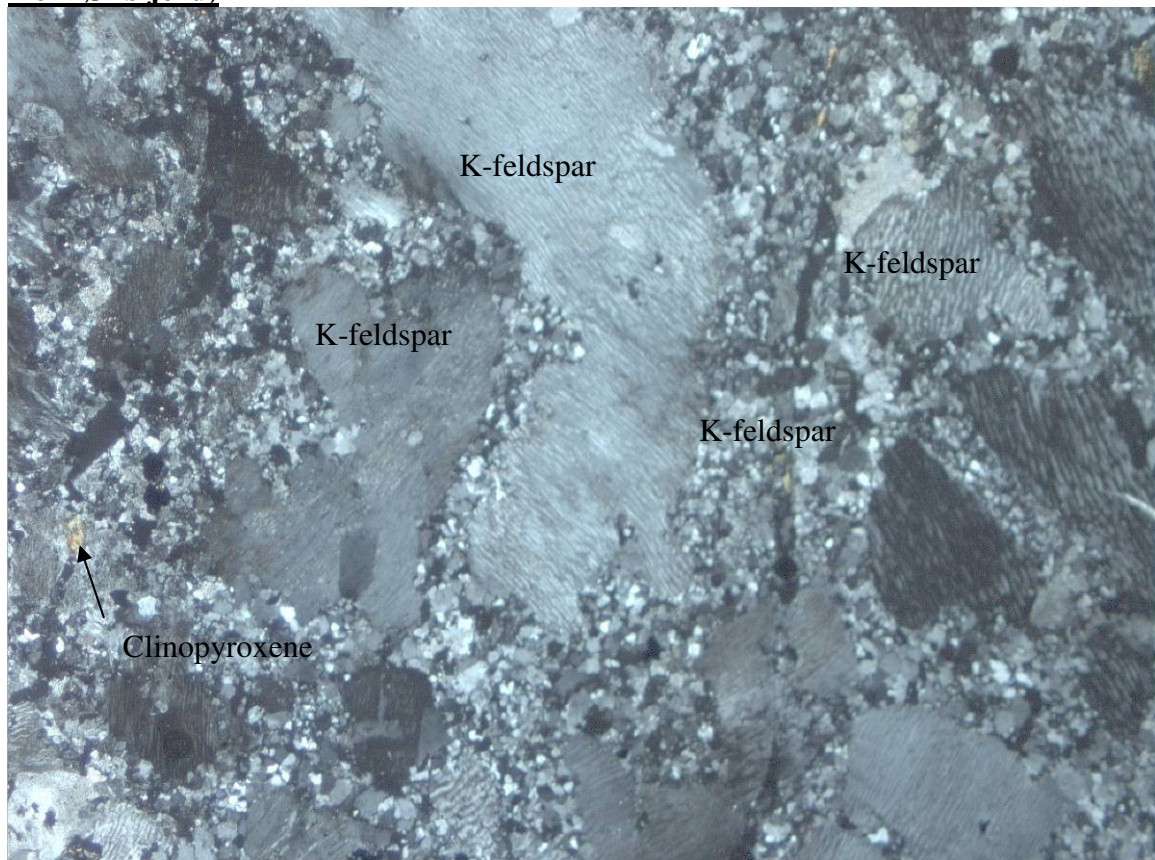
Figure A.16- Photomicrograph of sample RJR-02-41C (Monzodiorite from Øksfjord)



(Field of view 10mm. Cross-polarised light)

This rock consists of plagioclase (55%), K-feldspar (40%), and some quartz (5%). The feldspars occur as large anhedral phenocrysts, with agglomerations of recrystallised feldspar and quartz found at some of the grain boundaries, and little of the original igneous texture is preserved.

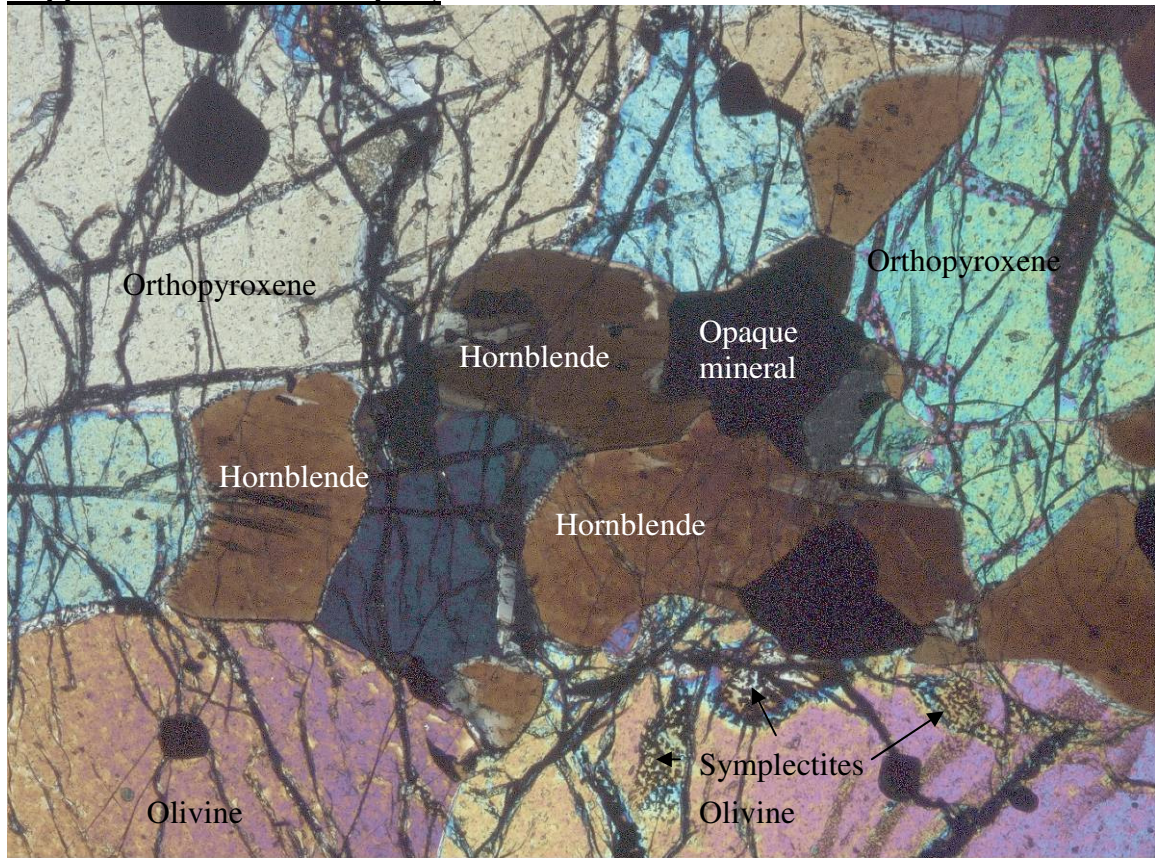
Figure A.17- Photomicrograph of sample RJR-03-101E (Alkali Feldspar Syenite from Øksfjord)



(Field of view 10mm. Cross-polarised light)

This rock is almost entirely composed of K-feldspar (95%), with small amounts of clinopyroxene (1%), biotite (2%), opaques (1%) and epidote (1%). The feldspar is found as both large anhedral crystals and as fine grained recrystallised material along the boundaries of the larger crystals.

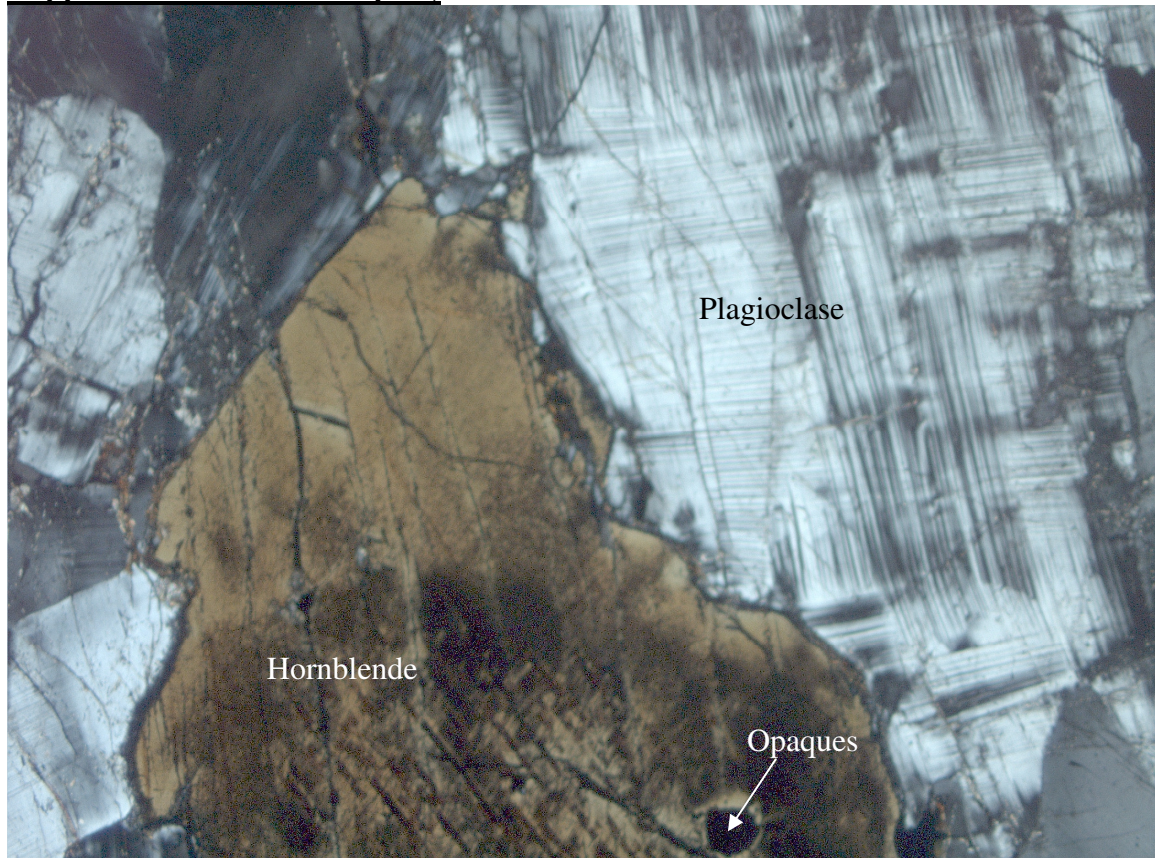
Figure A.18- Photomicrograph of sample RJR-03-108A (Harzburgite from Tappeluft Ultramafic Complex)



(Field of view 5mm. Cross-polarised light)

This coarse-grained cumulus rock consists primarily of olivine (40%), with orthopyroxene (10%), plagioclase (10%) and opaque minerals (5%). The remainder of the rock consists of metamorphic hornblende (30%) developed in the intercumulus spaces between the igneous crystals, and symplectic overgrowths of orthopyroxene and opaque spinel (5%).

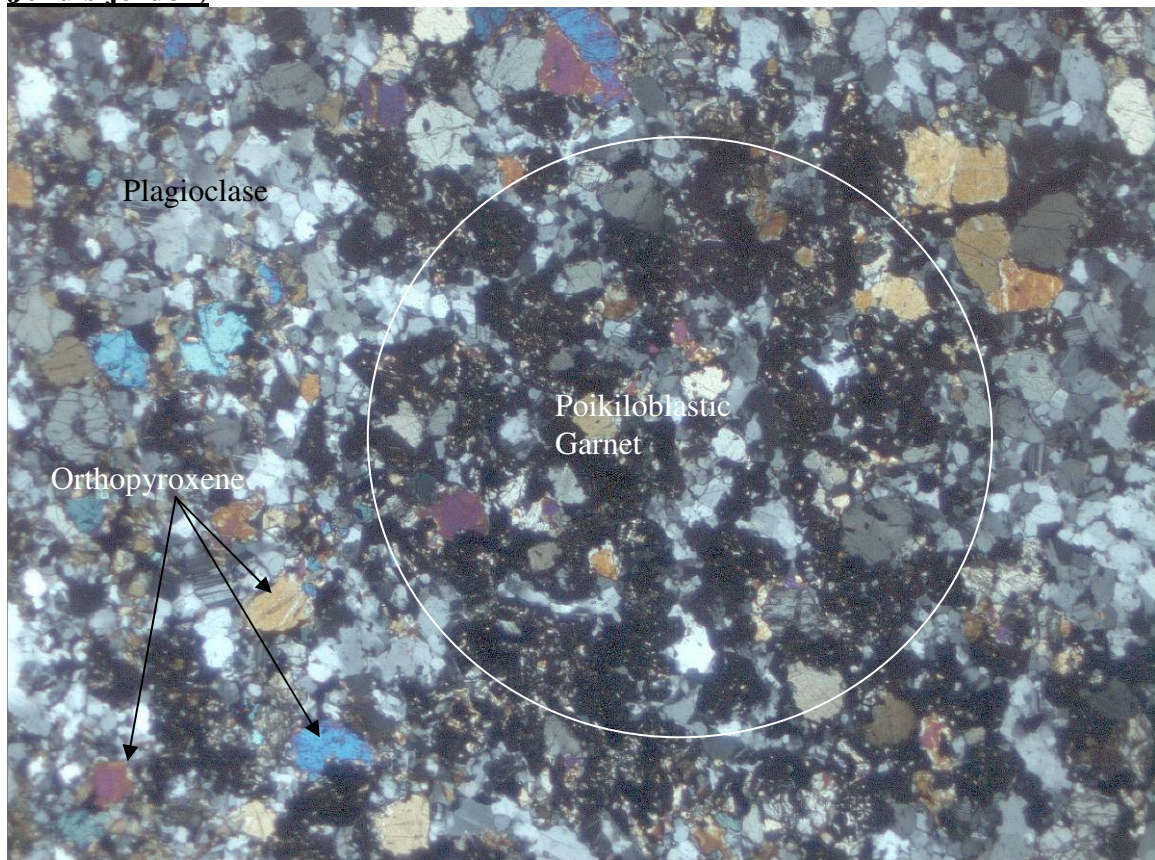
Figure A.19- Photomicrograph of sample RJR-03-109 (Hornblende gabbro from Tappeluft Ultramafic Complex)



(Field of view 10mm. Cross-polarised light)

This coarse grained rock is composed of hornblende (40%), plagioclase (40%), biotite (10%) and opaque minerals (10%). The plagioclase present is recrystallised, and hornblende often contains ilmenite exsolution lamellae more characteristic of clinopyroxene, the mineral which has been replaced by hornblende.

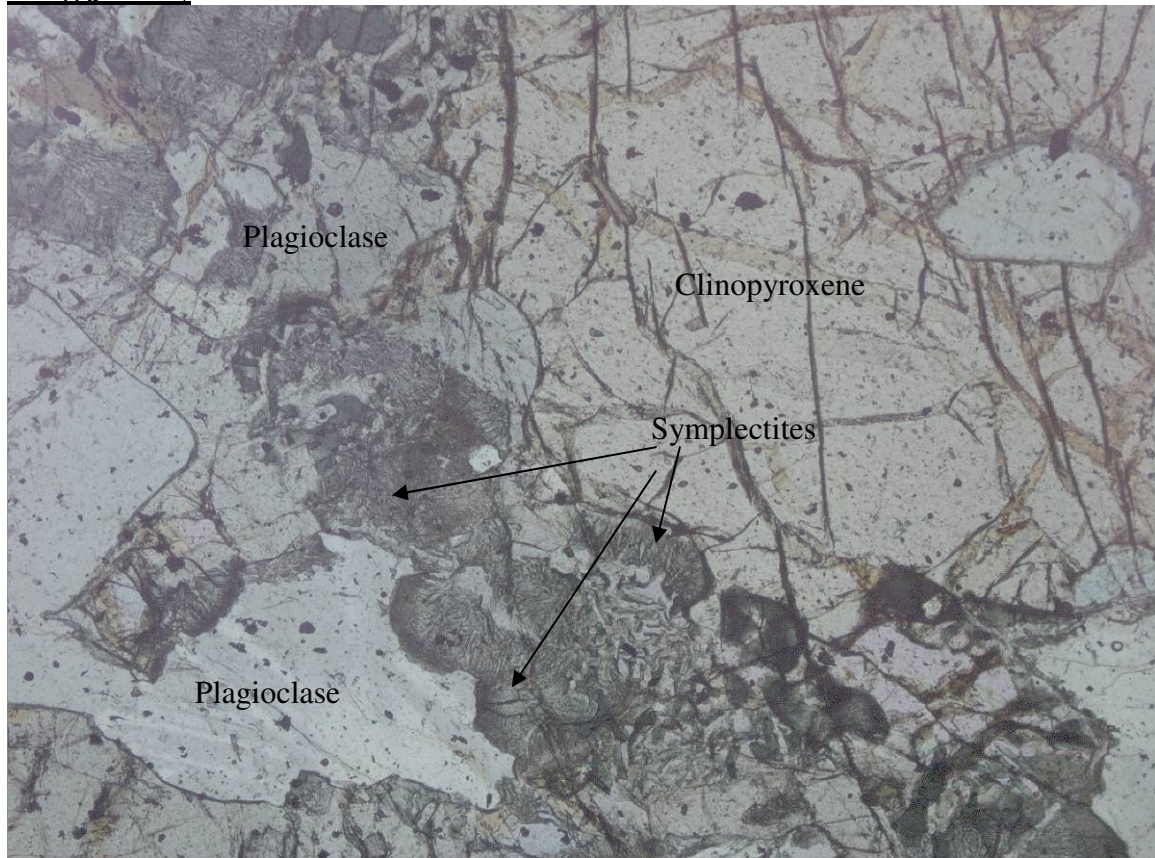
Figure A.20- Photomicrograph of sample RJR-03-112 (Metagabbro from Jokulfjorden)



(Field of view 10mm. Cross-polarised light)

This is a garnet-bearing metagabbro, which features the growth of large poikiloblastic garnets (5%) which include recrystallised plagioclase (30%), remnants of orthopyroxene (30%), opaque minerals (10%). Metamorphic hornblende (25%) is present as an overgrowth on many minerals.

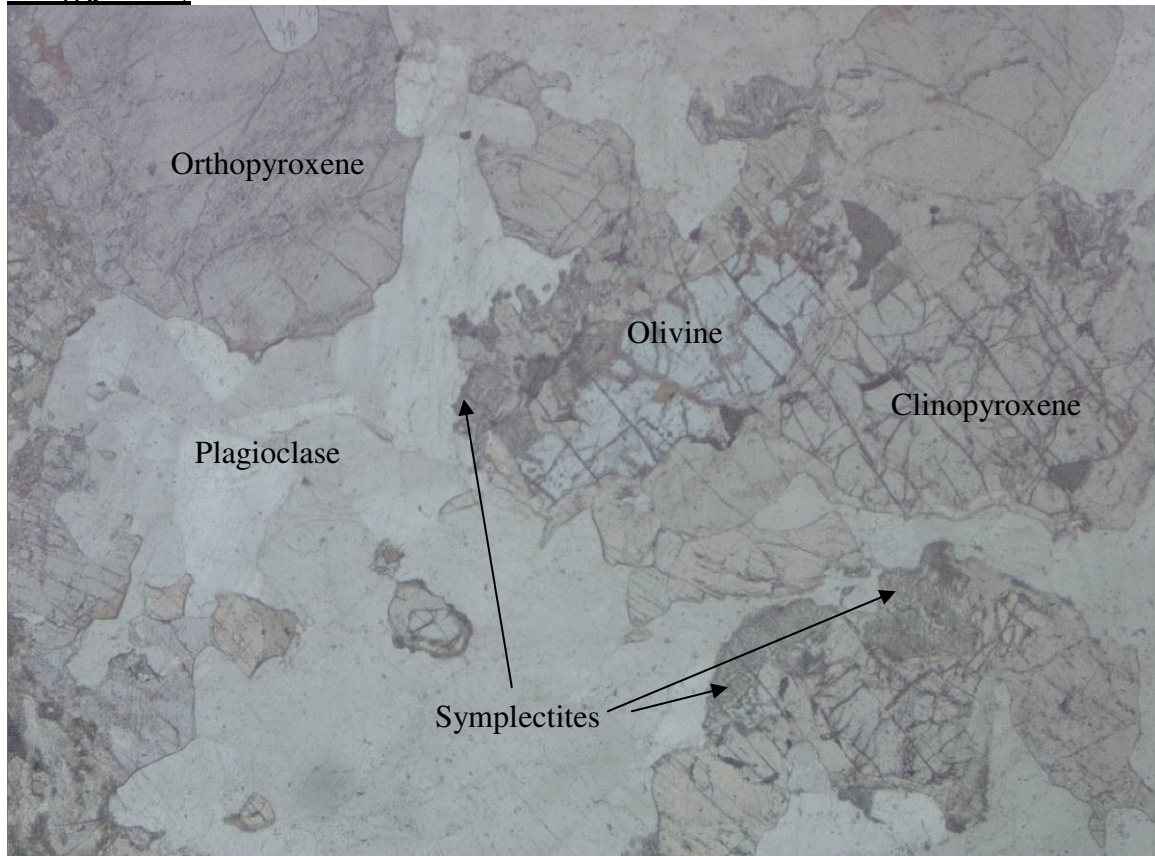
Figure A.21- Photomicrograph of sample RJR-03-113A (Metagabbro from Langfjorden)



(Field of view 5mm. Plane-polarised light)

This rock displays spectacular symplectic intergrowths of two different sorts. Orthopyroxene is intergrown with green spinel, and fayalitic olivine is intergrown with orthopyroxene. 10% of the rock consists of these intergrowths, with recrystallised plagioclase (50%), olivine (5%), orthopyroxene (10%) and clinopyroxene (15%) making up the bulk of the mineralogy. All crystals are heavily embayed except for the plagioclase, which shows clear recrystallisation features such as well-developed triple junctions.

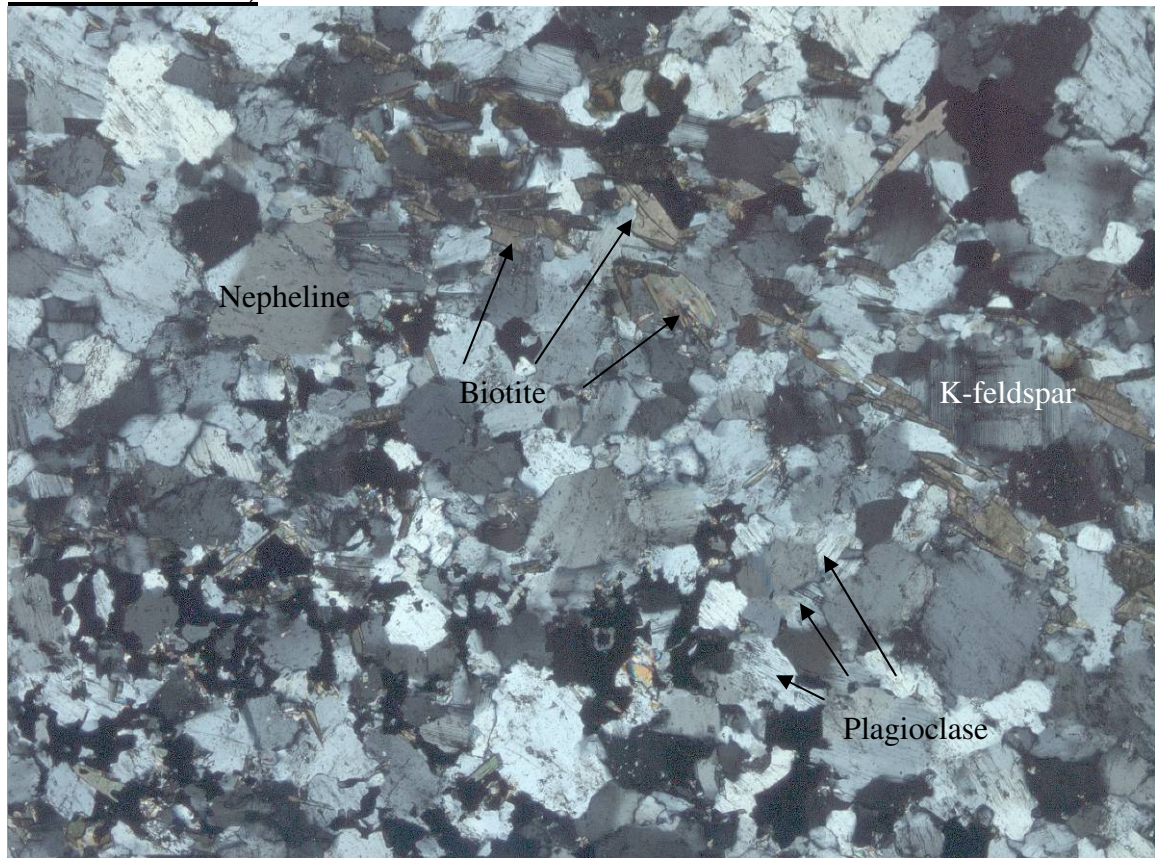
Figure A.22- Photomicrograph of sample RJR-03-113B (Gabbro from Langfjorden)



(Field of view 10mm. Plane-polarised light)

This gabbro has been metamorphosed to amphibolite grade, and contains a large amount (10%) of symplectic intergrowths of orthopyroxene and green spinel. Recrystallised plagioclase (50%) is abundant, and a small amount of metamorphic biotite (5%) is present as well. However, some of the original igneous mineralogy does remain, and clinopyroxene (15%), orthopyroxene (10%) and olivine (15%) are present as relict crystals.

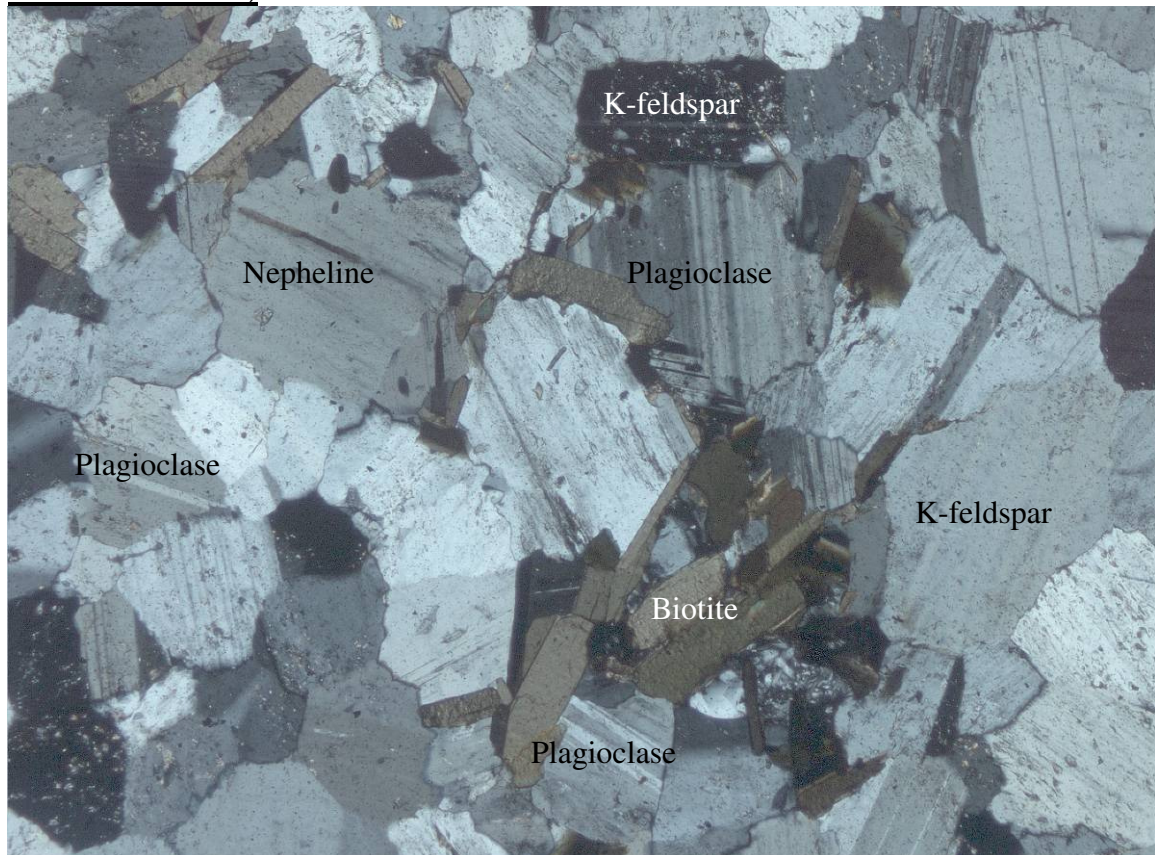
Figure A.23- Photomicrograph of sample RJR-03-115 (Nepheline Monzosyenite from Breivikbotn)



(Field of view 5 mm. Cross-polarised light)

This rock consists of plagioclase (20%), K-feldspar (40%), nepheline (20%), green biotite (10%), muscovite (5%), hornblende (3%) and opaques (2%). A foliation is described by elongate biotite crystals. Nepheline is present as large low relief equigranular crystals in amongst the recrystallised feldspars.

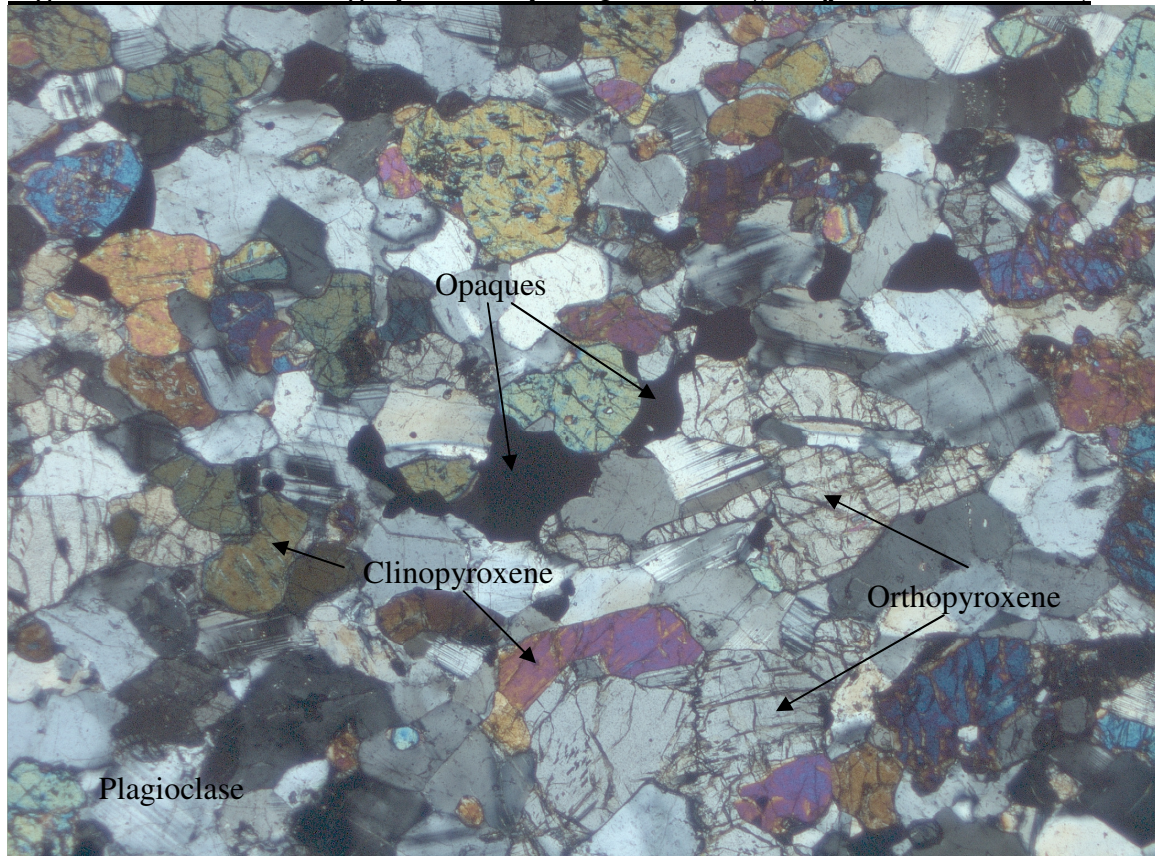
Figure A.24- Photomicrograph of sample RJR-03-116 (Nepheline Monzosyenite from Breivikbotn)



(Field of view 5 mm. Cross-polarised light)

This rock consists of plagioclase (20%), K-feldspar (30%), nepheline (30%), green biotite (10%), muscovite (5%), hornblende (3%) and opaques (2%). A foliation is described by elongate biotite crystals. Nepheline is present as large low relief equigranular crystals in amongst the recrystallised feldspars.

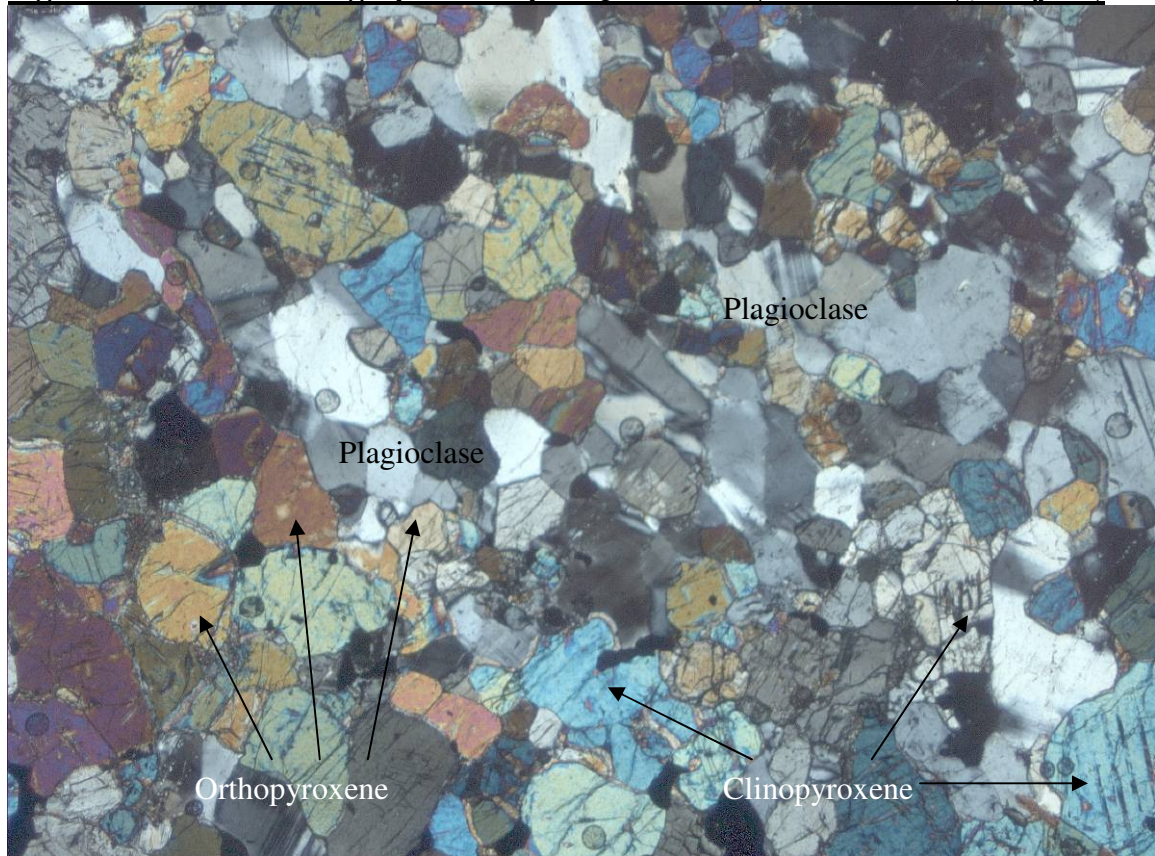
Figure A.25- Photomicrograph of sample RJR-03-118 (Øksfjord Gabbronorite)



(Field of view 10mm. Cross-polarised light)

This gabbronorite comprises recrystallised plagioclase (50%), orthopyroxene (20%), clinopyroxene (25%) and opaque minerals (5%). Much of the original igneous mineral assemblage is preserved, but the rock has been metamorphosed, as seen in the recrystallisation of the plagioclase grains.

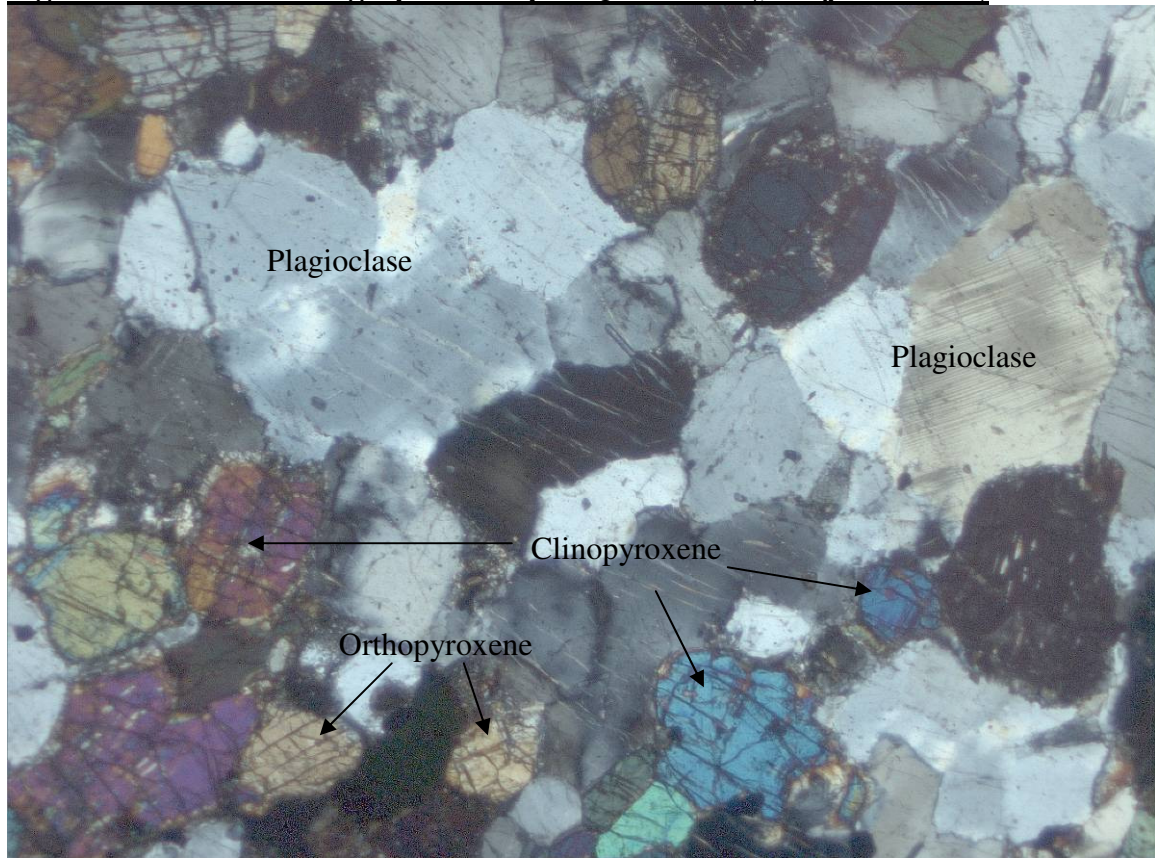
Figure A.26- Photomicrograph of sample R,JR-3-118B (Gabbronorite, Øksfjord)



(Field of view 10mm. Cross-polarised light)

This gabbronorite comprises plagioclase (50%), orthopyroxene (23%), clinopyroxene (25%), and minor opaques (2%). All minerals show the effects of metamorphism at amphibolite facies- plagioclase is recrystallised, and the edges of pyroxenes in contact with plagioclase are partly resorbed. However, many of the pyroxene crystals show good crystal habit, with many crystal faces being well preserved. The rock is an orthocumulate.

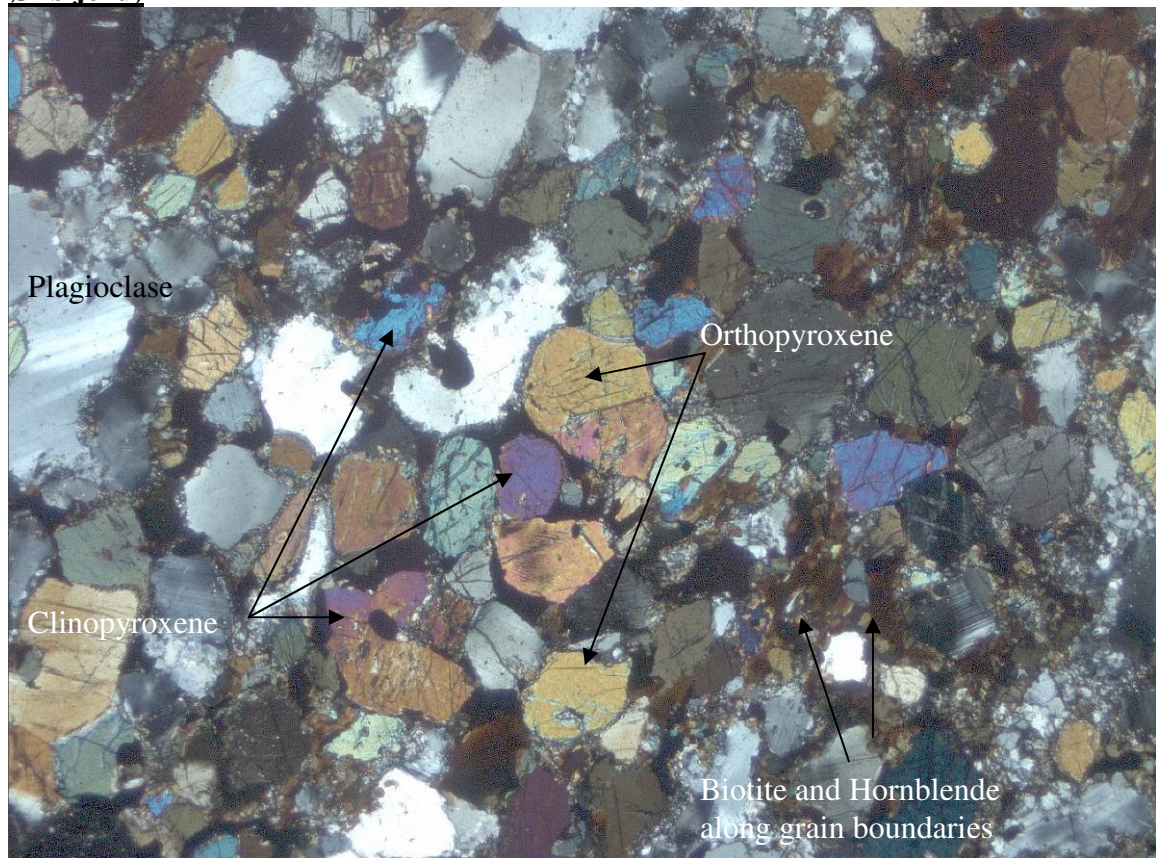
Figure A.27- Photomicrograph of sample RJR-03-120 (Øksfjord Norite)



(Field of view 10mm. Cross-polarised light)

This rock consists primarily of plagioclase (50%), which shows well-developed triple junctions indicative of recrystallisation. Orthopyroxene (30%) and opaques minerals (5%) are preserved from the original igneous assemblage, but a significant amount of metamorphic hornblende (15%) has been formed in this rock. The rock is medium-grained, and all crystals are anhedral.

Figure A.28- Photomicrograph of sample RJR-03-125 (Gabbronorite from Øksfjord)



(Field of view 10mm. Cross-polarised light)

This gabbronorite shows only limited growth of metamorphic hornblende (3%) and biotite (2%), both minerals generally growing along grain boundaries and in the intercumulus spaces in bands across the rock. The bulk of the rock is comprised of recrystallised plagioclase (30%), orthopyroxene (30%), clinopyroxene (30%) and opaque minerals (5%). The rock displays a foliation in hand specimen, created by the linear growth of the metamorphic minerals.

Figure A.29- Photomicrograph of sample RJR-03-129A (Granite from Øksfjord)



(Field of view 10mm. Cross-polarised light)

This foliated granite is composed of quartz (45%), k-feldspar (35%), and plagioclase (15%), along with some opaque minerals (5%). The crystals are generally elongated in the direction of the foliation as a result of metamorphic recrystallisation.

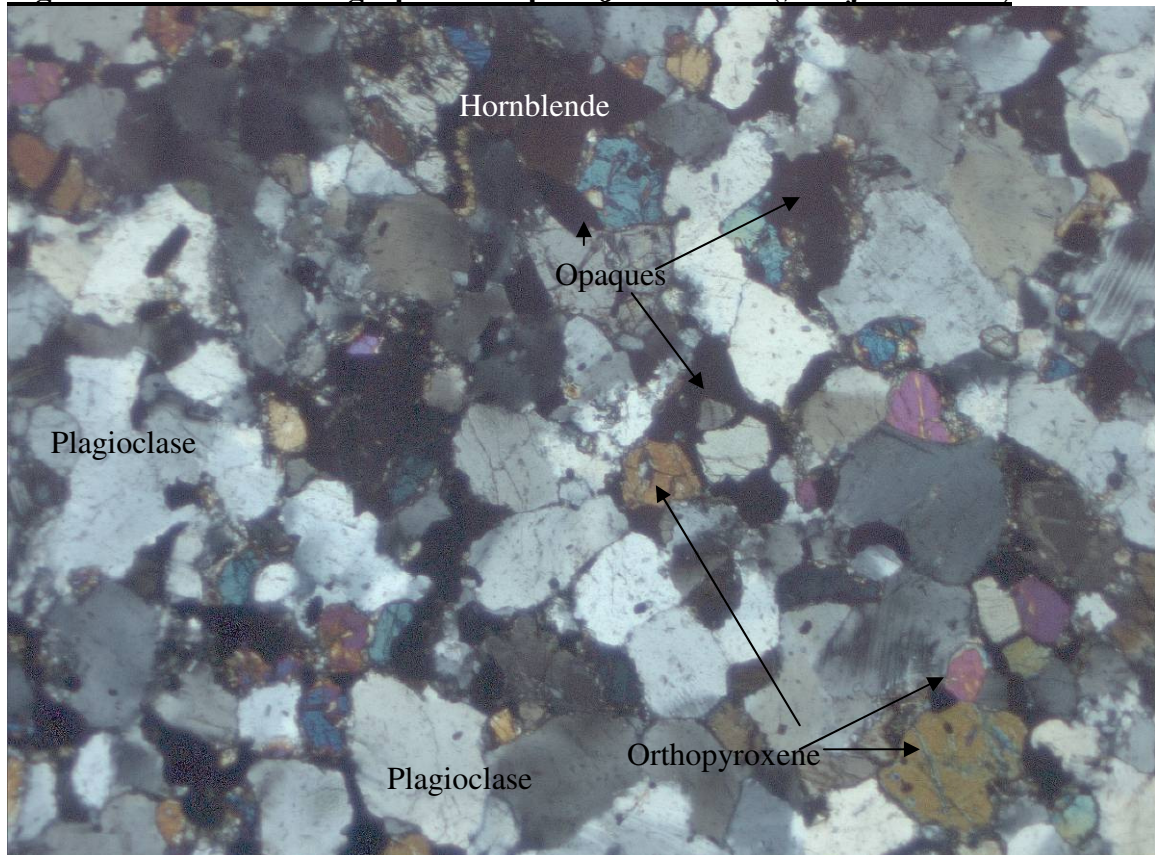
Figure A.30- Photomicrograph of sample RJR-03-129B (Granite from Øksfjord)



(Field of view 10mm. Cross-polarised light)

This granite has a distinct foliation, formed by both biotite and recrystallised feldspar. In composition, the rock consists of quartz (45%), plagioclase (20%), k-feldspar (25%), orthopyroxene (5%) and opaque minerals (5%). Zircon is present as an accessory phase.

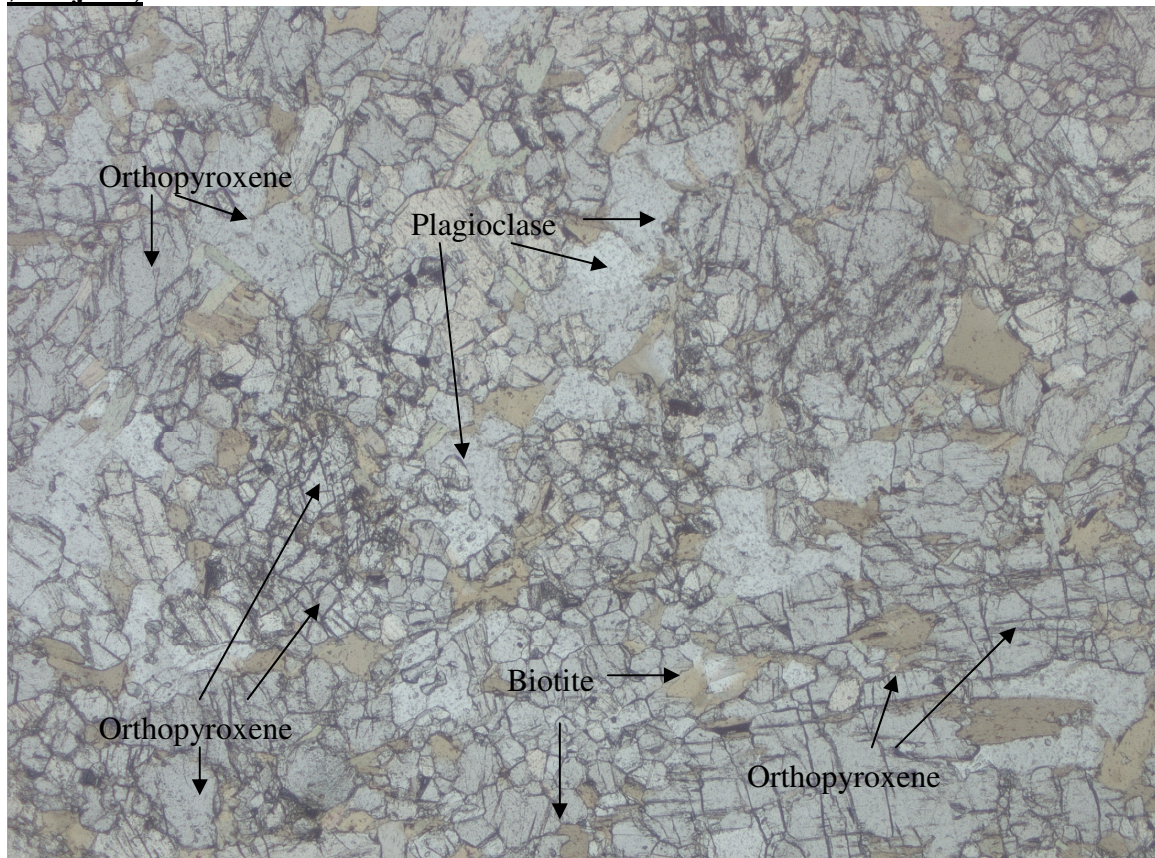
Figure A.31- Photomicrograph of sample RJR-03-129C (Øksfjord Norite)



(Field of view 10mm. Cross-polarised light)

This rock consists primarily of plagioclase (50%), which shows well-developed triple junctions indicative of recrystallisation. Orthopyroxene (30%) and opaques minerals (5%) are preserved from the original igneous assemblage, but a significant amount of metamorphic hornblende (15%) has been formed in this rock. The rock is medium-grained, and all crystals are anhedral.

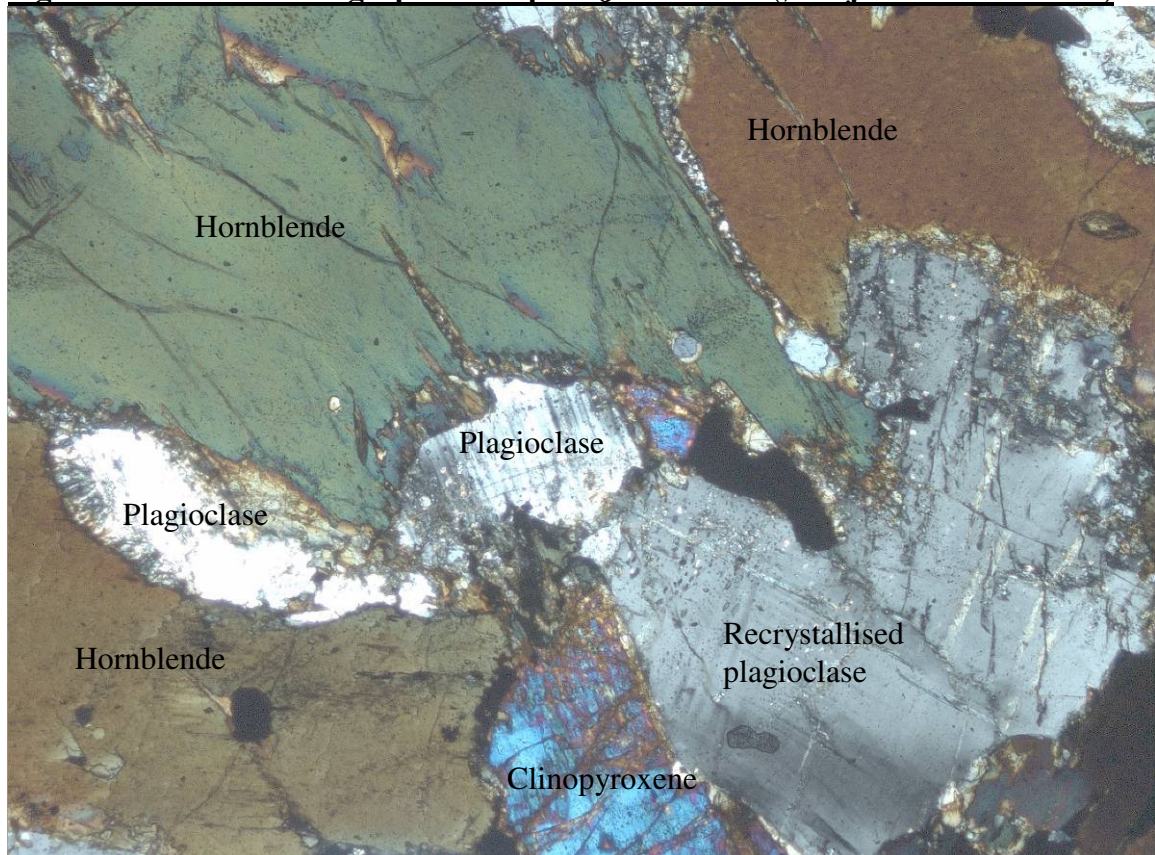
Figure A.32- Photomicrograph of sample RJR-03-129D (Orthopyroxenite from Øksfjord)



(Field of view 10mm. Plane-polarised light)

This rock is composed of orthopyroxene (70%), with some plagioclase (20%), biotite (5%) and opaque minerals (5%). The plagioclase is recrystallised, and biotite is growing on the edges of the orthopyroxenes.

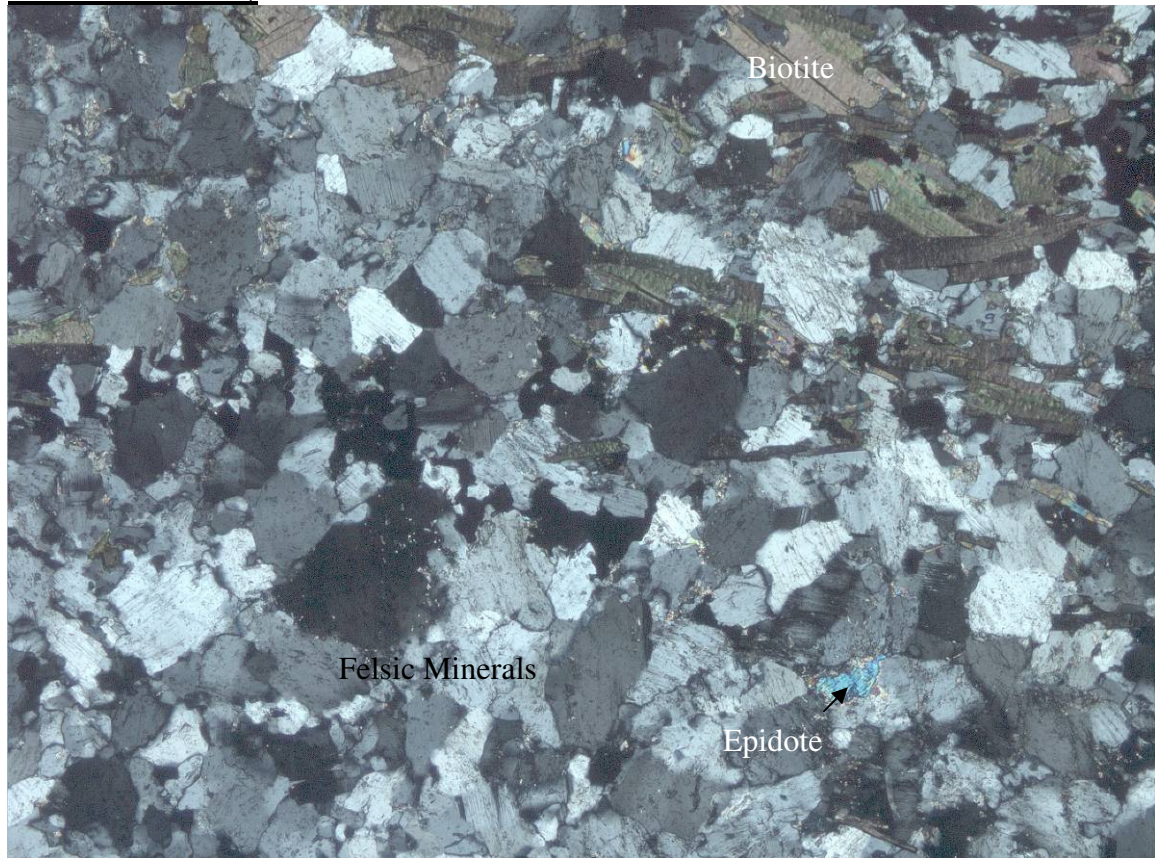
Figure A.33- Photomicrograph of sample RJR-03-130A (Øksfjord Gabbonorite)



(Field of view 5 mm. Cross-polarised light)

This coarse-grained rock has been extensively recrystallised, and recrystallised plagioclase (35%) and metamorphic hornblende (30%) dominate the section. Orthopyroxene (10%) and clinopyroxene (10%) are present, as well as opaque minerals (10%). Apatite and zircon were observed as accessory minerals. The pyroxenes are heavily resorbed, and little of the original igneous texture remains.

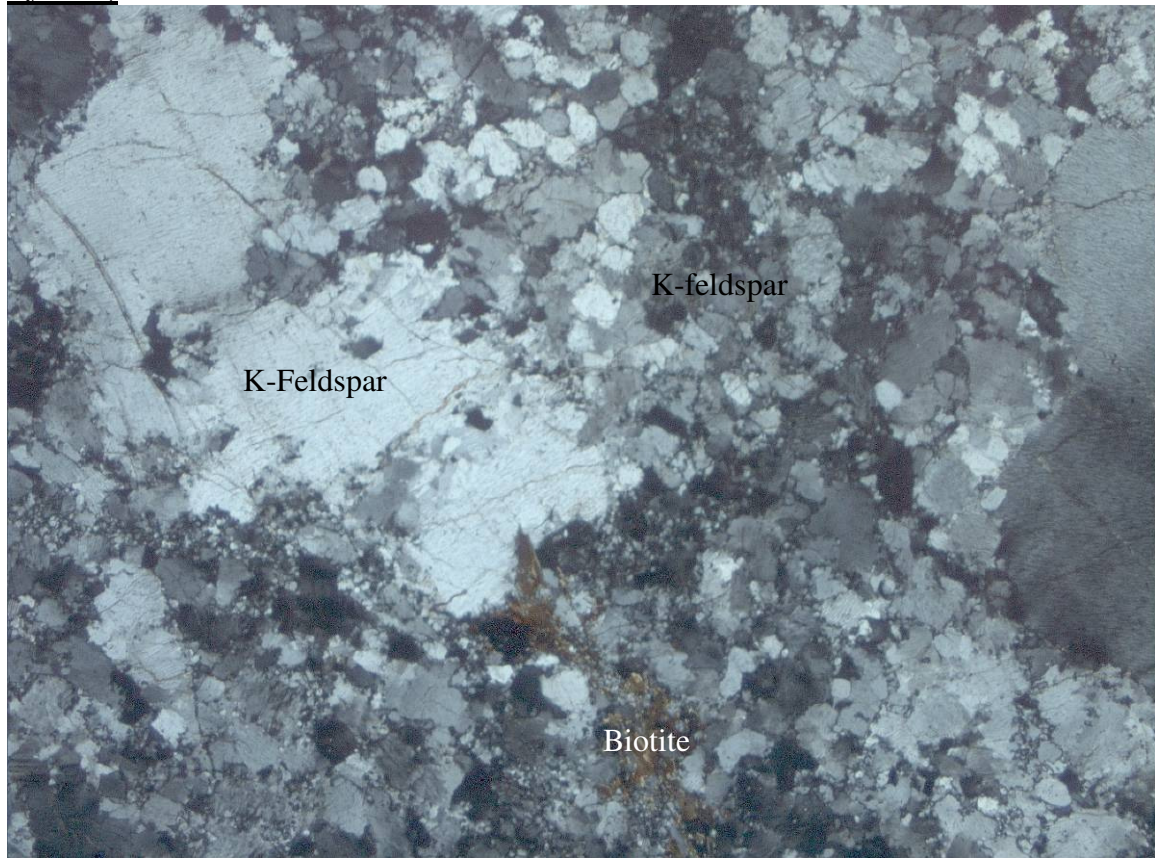
Figure A.34- Photomicrograph of sample RJR-04-236 (Nepheline Monzosyenite from Breivikbotn)



(Field of view 10mm. Cross-polarised light)

This rock consists of plagioclase (20%), K-feldspar (40%), nepheline (20%), green biotite (10%), hornblende (5%), epidote (3%) and opaques (2%). A foliation is described by elongate biotite crystals. Nepheline is present as large low relief equigranular crystals in amongst the recrystallised feldspars.

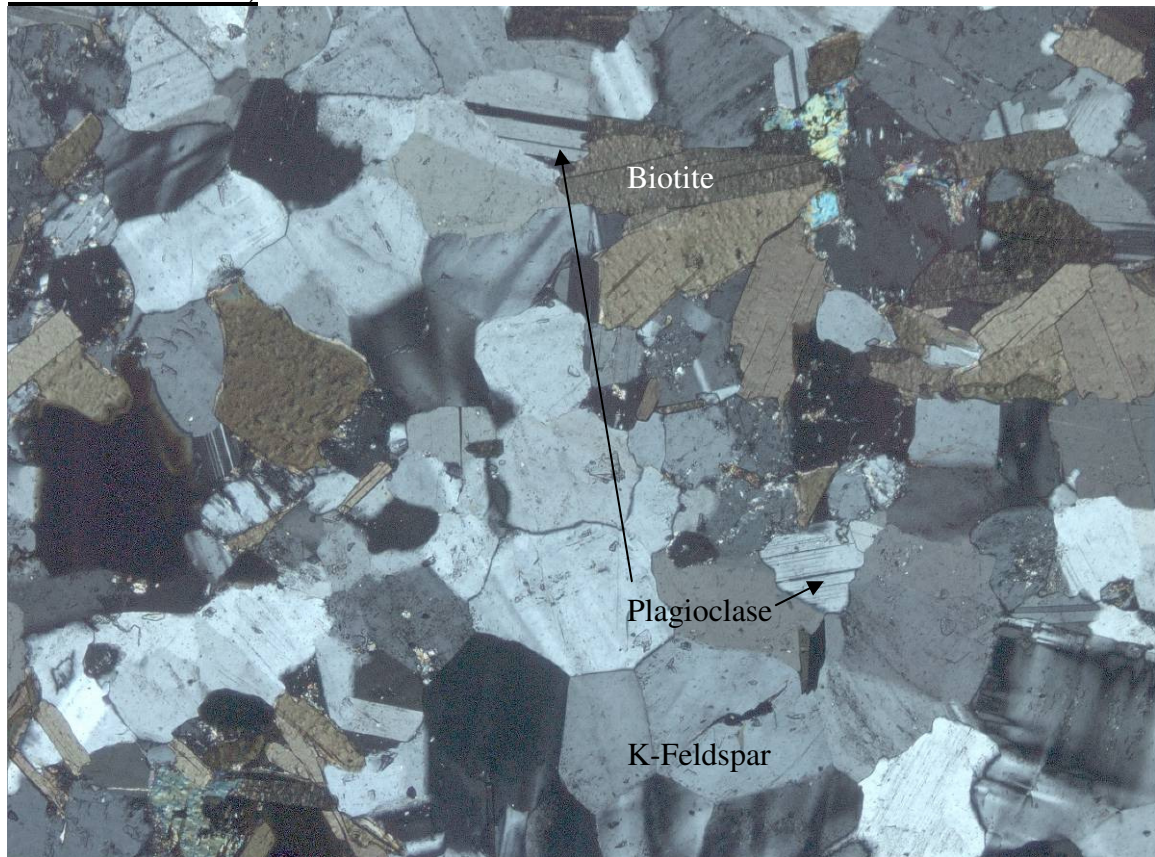
Figure A.35- Photomicrograph of sample RJR-04-245 (Nepheline-bearing Syenite)



(Field of view 10mm. Cross-polarised light)

This rock consists primarily of K-feldspar (90%), with subsidiary biotite (2%), opaque minerals (4%) and nepheline (4%). The rock has a metamorphic fabric defined by thin bands of elongated biotite. The K-feldspar crystals are anhedral and vary greatly in size, and display textures indicative of recrystallisation.

Figure A.36- Photomicrograph of sample RJR-04-246 (Nepheline Monzosyenite from Breivikbotn)



(Field of view 10mm. Cross-polarised light)

This rock consists of plagioclase (25%), K-feldspar (40%), nepheline (25%), green biotite (8%), and opaques (2%). A foliation is described by elongate biotite crystals. Nepheline is present as large low relief equigranular crystals in amongst the recrystallised feldspars. Zircon is present as an accessory mineral.

Appendix B:

Data for the Hasvik Gabbro

Received from Christian Tegner

Originally available as an electronic supplement

to:

**Tegner, C., Robins, B., Reginiussen, H., & Grundvig, S., 1999,
Assimilation of crustal xenoliths in a basaltic magma chamber: Sr
and Nd isotopic constraints from the Hasvik Layered Intrusion,
Norway, *Journal of Petrology* 40, 363- 380**

Table B.1 (background dataset): X-ray fluorescence analyses of whole rock

Sample#	CT1	CT2	CT3	CT4	CT5	P337"	CT6	CT7	CT8	CT9	CT10	CT11	CT12	CT13	CT14	CT15	CT16	CT17
Rock type	m.sed.	m.sed.	m.sed.	m.sed.	chill	chill	cumul.	cum.	cum.	cum.	cum.	cum.	cum.	cum.	cum.	cum.	cum.	cum.
Location/Zone	Aureole	Aureole	Aureole	Aureole	MBS	MBS	BZ	BZ	BZ	BZ	BZ	LZ	LZ	LZ	LZ	LZ	LZ	LZ
Strat. pos. (m)	5*	50*	100*	45*	-	-	22	50	65	82	107	117	163	209	249	297	335	377
<i>Major element (wt%)</i>																		
SiO ₂	47.04	64.52	60.76	87.89	49.61	49.57	50.74	50.41	51.59	50.82	52.51	49.14	46.22	49.61	49.38	50.63	49.06	50.09
TiO ₂	3.43	0.72	1.03	0.17	1.79	1.02	0.61	0.64	0.50	0.47	0.47	0.48	0.37	0.49	0.51	0.63	0.41	0.53
Al ₂ O ₃	23.96	16.59	16.98	6.37	15.01	15.55	18.01	18.15	19.86	17.45	18.76	17.51	13.26	17.94	16.60	14.06	21.33	17.38
FeO	5.78	3.81	5.78	1.25	6.69	7.15	3.91	4.21	4.08	4.25	3.23	3.50	6.12	3.36	4.18	3.64	2.89	3.50
Fe ₂ O ₃	2.41	1.43	0.93	0.32	3.05	0.79	3.64	2.45	1.44	2.57	2.62	1.99	3.01	2.01	1.85	2.81	1.62	2.38
MnO	0.11	0.21	0.14	0.10	0.16	0.14	0.12	0.11	0.09	0.12	0.10	0.10	0.13	0.10	0.11	0.12	0.08	0.10
MgO	3.97	1.51	1.63	0.43	7.63	9.37	7.64	7.53	7.01	8.09	7.86	7.89	14.51	7.57	8.70	9.37	6.54	8.15
CaO	9.45	5.14	1.70	1.61	11.63	12.87	11.37	11.98	12.85	11.40	12.78	14.98	11.66	14.74	15.40	15.78	15.01	14.04
Na ₂ O	2.58	2.58	2.89	b.d.	2.04	2.10	2.07	1.99	2.37	1.99	2.34	1.61	1.31	1.65	1.51	1.39	1.74	1.95
K ₂ O	0.24	0.69	5.42	1.60	0.37	0.27	0.21	0.18	0.33	0.22	0.16	0.13	0.07	0.09	0.13	0.05	0.10	0.13
P ₂ O ₅	0.03	0.05	0.25	b.d.	0.09	0.14	0.04	0.01	0.02	0.01	0.03	0.02	0.03	0.01	0.02	0.02	0.03	0.04
L.O.I	0.83	1.20	0.76	0.23	0.53	0.47	0.28	0.53	0.69	0.86	0.29	0.89	0.93	1.13	0.39	0.26	0.66	0.73
Sum	99.83	98.45	98.27	99.97	98.60	99.44	98.64	98.19	100.83	98.25	101.15	98.24	97.62	98.70	98.78	98.76	99.47	99.02
<i>Whole rock composition</i>																		
mg#	0.471	0.346	0.305		0.591	0.680	0.655	0.677	0.699	0.688	0.715	0.727	0.746	0.723	0.726	0.731	0.728	0.720
<i>Trace element (ppm)</i>																		
V	747	121	111	7	297		149	156	157	124	136	142	87	137	161	180	96	143
Cr	183	77	76	21	130	320	133	165	272	187	171	415	503	425	694	453	308	334
Ni	69	25	27	9	35	552#	39	71	59	51	36	91	231	62	115	53	61	33
Cu	30	23	b.d.	10	16	12#	21	57	40	11	23	68	18	42	75	34	39	16
Zn	124	59	112	6	77	63#	49	36	41	49	29	29	46	25	34	24	12	27
Y	10	25	65	12	15	7#	9	9	10	9	9	9	9	9	9	12	8	9
Zr	70	193	464	104	62	39#	31	27	25	25	23	24	24	24	24	24	23	30
Nb	13	12	29	4	8	4	3	2	3	3	3	b.d.	3	b.d.	b.d.	b.d.	2	3
Ba	253	530	1580	339	187	110	130	60	122	73	94	57	84	103	82	45	b.d.	57
Ce	27	57	47	25	24	14#	14	15	10	13	17	21	18	10	27	20	21	21

Table B.1 Continued

Sample#	CT18	CT19	CT20	CT21	CT22	CT23	CT24	CT25	CT26	CT27	CT28	CT29	CT30	CT31	CT32	CT33	CT34	CT35	CT40	CT41	CT43	CT44
Rock type	cum.	cum.	cum.	cum.	cum.	cum.	cum.	cum.	cum.	cum.	cum.	cum.	cum.	cum.	cum.	cum.	cum.	cum.	cum.	cum.	cum.	cum.
Location/Zone	MZa	MZa	MZa	MZa	MZa	MZa	MZa	MZa	MZa	MZb	MZb	MZb	MZb	MZb	MZb	MZb	MZb	MZb	UZ	UZ	UBS	UBS
Strat. pos. (m)	444	505	530	564	600	648	692	747	789	843	907	959	1013	1066	1125	1181	1226	1273	1397	1447	1551	1568
<i>Major element (wt%)</i>																						
SiO ₂	51.43	52.24	51.41	52.73	51.15	51.54	51.67	51.19	52.84	46.21	45.85	46.01	45.88	50.06	49.62	46.90	46.50	47.99	46.38	46.29	47.82	47.34
TiO ₂	0.58	0.49	0.50	0.43	0.58	0.51	0.44	0.43	0.47	6.73	6.95	6.60	6.00	2.81	1.33	5.81	4.88	5.25	3.78	3.32	2.97	3.31
Al ₂ O ₃	13.37	16.14	18.43	19.56	16.47	19.46	20.45	18.85	18.37	15.93	15.97	15.66	14.89	16.67	9.76	15.77	15.43	16.12	15.24	15.07	15.83	15.69
FeO	4.14	4.69	3.40	3.50	3.77	3.53	3.80	3.87	3.60	8.39	7.78	7.54	6.49	5.98	5.91	7.88	7.17	7.61	11.42	11.66	8.67	8.39
Fe ₂ O ₃	3.64	2.22	2.88	2.90	3.71	3.14	2.03	2.59	3.58	3.67	4.72	4.98	6.29	3.52	5.80	4.60	5.26	4.77	3.37	4.61	5.13	6.18
MnO	0.15	0.12	0.11	0.11	0.13	0.11	0.11	0.12	0.13	0.15	0.15	0.16	0.17	0.15	0.21	0.17	0.17	0.18	0.24	0.26	0.17	0.18
MgO	10.36	9.49	7.67	7.97	8.80	7.53	6.68	6.82	7.47	7.10	6.86	6.76	7.00	6.68	9.20	6.58	6.41	6.15	4.60	3.86	6.09	6.23
CaO	12.65	11.81	11.79	10.39	11.58	10.27	10.38	11.17	10.73	10.68	10.65	10.55	10.51	11.58	15.72	10.22	9.75	9.86	9.87	9.39	9.72	9.42
Na ₂ O	2.24	2.36	2.14	2.51	1.96	2.35	2.54	2.50	2.44	2.00	2.00	2.04	2.07	2.19	1.28	2.18	2.30	2.49	2.59	2.61	2.54	2.58
K ₂ O	0.13	0.27	0.16	0.19	0.14	0.17	0.19	0.21	0.17	0.12	0.12	0.13	0.14	0.14	0.06	0.15	0.19	0.18	0.19	0.28	0.46	0.50
P ₂ O ₅	0.03	0.02	0.03	0.03	0.03	0.02	0.02	0.02	0.02	0.02	0.01		0.02		0.03	0.02	0.03	0.02	2.50	2.36	0.31	0.27
L.O.I	0.33	0.66	0.36	0.46	0.30	0.23	0.38	0.63	0.10	0.00	0.00	0.00	0.00	0.20	0.23	0.00	0.00	0.00	0.00	0.00	0.66	0.00
Sum	99.05	100.51	98.88	100.78	98.62	98.86	98.69	98.40	99.92	101.00	101.06	100.43	99.46	99.98	99.15	100.28	98.09	100.62	100.18	99.71	100.37	100.09
<i>Whole rock composition</i>																						
mg#	0.714	0.717	0.695	0.699	0.688	0.679	0.679	0.662	0.661	0.520	0.504	0.501	0.507	0.566	0.596	0.494	0.490	0.480	0.362	0.303	0.450	0.443
<i>Trace element (ppm)</i>																						
V	149	157	130	112	138	111	109	138	137	638	708	666	603	385	449	639	597	632	626	466	418	427
Cr	286	211	154	118	147	96	78	90	85	98	56	47	84	88	682	74	43	59	35	34	177	146
Ni	40	48	36	33	31	28	30	28	28	33	30	20	22	18	88	27	15	26	11	11	104	84
Cu	9	22	16	15	14	14	18	23	14	22	22	19	16	13	24	22	15	14	9	11	43	37
Zn	33	47	32	37	38	38	37	42	40	37	53	40	48	43	62	55	63	61	105	121	98	101
Y	10	9	8	7	10	7	7	6	7	8	8	10	10	10	17	9	9	10	30	34	19	17
Zr	26	23	28	28	28	26	25	22	23	32	33	36	37	25	30	40	38	42	39	46	85	82
Nb	3	3	2	3	3	2	3	b.d.	b.d.	7	8	9	9	5	3	11	12	13	18	19	19	20
Ba	73	88	60	136	48	93	104	b.d.	203	113	61	60	125	82	101	156	69	108	143	150	185	184
Ce	15	17	21	12	15	26	10	14	19	17	13	19	23	11	19	12	17	13	43	42	32	34

Table B.2 (background dataset): Sr and Nd Isotopic analyses of whole rock

Sample#	Location/ Zone	Strat. pos. (m)	Rb (ppm)	Sr (ppm)	⁸⁷ Rb/ ⁸⁶ Sr	⁸⁷ Sr/ ⁸⁶ Sr measured	⁸⁷ Sr/ ⁸⁶ Sr (700 Ma)	ε _{Sr} (700 Ma)	Sm (ppm)	Nd (ppm)	¹⁴⁷ Sm/ ¹⁴⁴ Nd	¹⁴³ Nd/ ¹⁴⁴ Nd measured	¹⁴³ Nd/ ¹⁴⁴ Nd (700 Ma)	ε _{Nd} (700 Ma)	
Mafic cumulate															
CT44	UBS	1568	9.97	652.5	0.0444	0.705819	##	0.705375	24.02 (48)	3.41	10.03	0.2068	0.512495 (6)	0.511546	-3.69 (12)
CT43	UBS	1551	8.32	518.3	0.0466	0.705649	##	0.705183	21.29 (114)	4.43	10.75	0.2510	0.512496 ##	0.511344	-7.63 ##
CT41	UZ	1447	5.02	533.7	0.0273	0.709179	##	0.708906	74.20 (40)	8.34	29.22	0.1739	0.512367 ##	0.511569	-3.24 ##
CT40	UZ	1397	0.74	514.0	0.0042	0.708695	##	0.708653	70.60 (55)	6.60	27.70	0.1440	0.512409 (7)	0.511748	0.26 (14)
CT35	MZb	1273	1.75	461.2	0.0110	0.707235	##	0.707125	48.89 (34)	1.01	3.60	0.1700	0.512615 (7)	0.511835	1.95 (14)
CT33	MZb	1181	1.65	494.3	0.0097	0.706856	(16)	0.706759	43.69 (23)	0.90	3.30	0.1710	0.512591 (6)	0.511806	1.39 (12)
CT31	MZb	1066	0.85	512.5	0.0048	0.706591	(16)	0.706543	40.62 (23)	1.06	3.10	0.2068	0.512663 (7)	0.511714	-0.41 (14)
CT29	MZb	959	1.22	459.9	0.0077	0.706209	(14)	0.706132	34.78 (20)	0.78	2.90	0.1770	0.512648 (8)	0.511836	1.97 (16)
CT27	MZb	843	2.67	400.5	0.0194	0.705993	##	0.705799	30.04 (31)	0.88	2.86	0.1853	0.512660 (8)	0.511810	1.46 (16)
CT25	MZa	747	3.42	521.1	0.0191	0.705362	##	0.705171	21.12 (37)	0.78	2.89	0.1636	0.512581 (8)	0.511830	1.86 (16)
CT23	MZa	648	3.22	522.8	0.0179	0.704849	##	0.704670	14.00 (57)	0.80	3.30	0.1471	0.512553 (8)	0.511878	2.80 (16)
CT21	MZa	564	2.33	511.6	0.0132	0.704864	##	0.704732	14.88 (28)	0.79	3.32	0.1443	0.512530 (5)	0.511868	2.60 (10)
CT19	MZa	505	11.12	458.7	0.0704	0.705135	(18)	0.704432	10.61 (26)	1.24	4.67	0.1611	0.512718 ##	0.511979	4.76 (51)
CT18	MZa	444	1.90	342.6	0.0161	0.704692	(16)	0.704531	12.02 (23)	1.21	3.98	0.1838	0.512710 (7)	0.511867	2.57 (14)
CT17	LZ	377	3.09	469.2	0.0195	0.704031	##	0.703836	2.15 (28)	1.29	4.76	0.1650	0.512703 (10)	0.511946	4.12 ##
CT16	LZ	335	2.04	517.7	0.0114	0.703910	(14)	0.703796	1.58 (20)	1.02	3.87	0.1604	0.512689 (12)	0.511953	4.26 ##
CT15	LZ	297	0.53	301.0	0.0048	0.704402	##	0.704354	9.51 (28)	1.52	4.86	0.1909	0.512748 (16)	0.511872	2.68 (31)
CT14	LZ	249	3.60	374.3	0.0279	0.704477	(16)	0.704198	7.29 (23)	1.23	4.05	0.1846	0.512731 (7)	0.511884	2.91 (14)
CT13	LZ	209	2.26	419.0	0.0156	0.704581	(18)	0.704425	10.52 (26)	1.20	4.10	0.1783	0.512697 (9)	0.511879	2.81 (18)
CT12	LZ	163	1.47	315.1	0.0135	0.704429	(18)	0.704294	8.66 (26)	0.83	2.87	0.1756	0.512717 (8)	0.511911	3.45 (16)
CT11	LZ	117	3.45	161.6	0.0619	0.704550	##	0.703932	3.51 (34)	1.17	4.13	0.1730	0.512691 (8)	0.511897	3.17 (16)
CT9	BZ	82	5.74	452.7	0.0368	0.705129	(16)	0.704761	15.30 (23)	0.98	3.71	0.1599	0.512562 (6)	0.511828	1.82 (12)
CT7	BZ	50	3.99	430.9	0.0250	0.704952	##	0.704702	14.46 (68)	1.08	3.91	0.1682	0.512612 (8)	0.511840	2.06 (16)
Mafic chill															
CT5	MBS	-	13.30	448.1	0.0860	0.705340	(15)	0.704481	11.31 (21)	2.50	10.20	0.1475	0.512567 (4)	0.511890	3.03 (8)
P337	MBS	-	3.75	173.5	0.0630	0.704526	(13)	0.703897	3.01 (18)	2.20	8.60	0.1563	0.512598 (6)	0.511881	2.85 (12)
Country rock															
CT1	Aureole	5*	3.19	562.9	0.0165	0.708586	##	0.708421	67.31 (28)	1.53	9.22	0.1002	0.512024 (13)	0.511564	-3.34 ##
CT2	Aureole	50*	12.37	368.3	0.0977	0.720813	##	0.719837	229.53 (34)	4.90	30.89	0.0965	0.511903 (14)	0.511460	-5.37 ##
CT3	Aureole	100*	220.53	251.3	2.5574	0.741432	(18)	0.715885	173.37 (26)	15.07	80.41	0.1141	0.511955 (7)	0.511431	-5.94 (14)
CT4	Aureole	45*	54.83	111.4	1.4276	0.732302	(18)	0.718041	204.01 (26)	2.61	15.51	0.1025	0.511878 (7)	0.511408	-6.39 (14)

Uncertainties reported on ⁸⁷Sr/⁸⁶Sr and ¹⁴³Nd/¹⁴⁴Nd are 2σ analytical errors of last significant decimals; uncertainties for ε-values are propagated 2σ analytical values.

⁸⁷Sr/⁸⁶Sr are normalised to ⁸⁶Sr/⁸⁸Sr = 0.1196 and ¹⁴³Nd/¹⁴⁴Nd to ¹⁴⁶Nd/¹⁴⁴Nd = 0.721903.

Sr calculations assumes the decay constant λ⁸⁷Rb = 1.42e⁻¹¹ yr⁻¹ and UR (700 Ma) ⁸⁷Sr/⁸⁶Sr = 0.703685; Nd calculations assumes the decay constant λ¹⁴⁷Sm = 6.54e⁻¹² yr⁻¹ and CHUR (700 Ma) ¹⁴³Nd/¹⁴⁴Nd = 0.51173.

